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The Loch Lomond Stadial ice cap in Western Lochaber, Scotland.

Debbie Greene

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Declaration

I declare that this thesis has been composed by myself and is entirely my own work, except where otherwise indicated.

Deborah R. Greene.

Abstract

The aim of this thesis is to reconstruct the dimensions of the Loch Lomond Stadial (LLS) ice cap in Western Lochaber, Scotland, and to make inferences about ice cap dynamics on the basis of the field evidence. There have been no previous detailed studies of LLS ice cover in Western Lochaber, yet an accurate empirical reconstruction of the ice cap is an essential input and constraint to models of the interactions between climate, ice volume, topography and mantle rheology.

Three different types of field evidence are used to reconstruct the ice limits. Firstly, geomorphological and sedimentary evidence indicates some lateral and terminal limits and retreat patterns. Secondly, a seismic stratigraphy of Loch Linnhe and the Firth of Lorne provides additional evidence for the distribution of glacial deposits and an ice limit. Thirdly, the glacial and periglacial evidence on 111 slopes marks trimline altitudes reflecting the palaeo-ice surface.

Ice limits reconstructed from these three types of evidence allow a three dimensional reconstruction of the ice surface. The ice cap was up to ~650m thick, and ice flowed from the main mountain ridges and an ice plateau around the heads of Lochs Eil and Shiel, down the main troughs to the sea lochs. All available chronological controls support the proposition that this ice cap existed during the LLS. Depositional evidence suggests that subsequent ice retreat back towards the mountains was punctuated by stillstands at topographic pinning points in the sea lochs. There are distinct contrasts in the spatial distributions of glacial erosional and depositional evidence. Throughout the west and south of Western Lochaber terrestrial signs of glacial scouring are widespread and glacial till and moraines are thin and sporadic, yet there are thick glacimarine sequences in the proglacial submarine basin in Loch Linnhe, and large proglacial outwash fans around the lochs. In the north east of Western Lochaber and in Eastern Lochaber slopes are mostly mantled with glacial till, there are large terminal and lateral moraines and outwash deposits are common.

These patterns may be explained mainly by a consideration of tidewater glacier theory and topographic controls on ice dynamics. In particular, the iceberg calving termini of the outlet glaciers may have resulted in high velocity ice flow and erosional regimes right to the snouts, so that deposition was concentrated offshore or in proglacial fluvioglacial outwash spreads. Trough geometry determined the sites of stillstands of the former tidewater glaciers during retreat. Several lines of evidence suggest that landscape modification during the LLS may have been relatively insubstantial. The reconstructed ice surface allows the estimation of palaeo-Equilibrium Line Altitudes. These are calculated using an AAR method, which allows for advancing and calving glacial regimes. The derived ELAs range from 309 - 478m, with a regional mean of 360m. This value accords with previous ELA reconstructions in the south west Highlands.

Acknowledgements

I am immensely grateful to David Sugden for inspirational encouragement, guidance and assistance throughout the last three years, and for wise comments on earlier drafts of this thesis. I also thank Geoffrey Boulton for invaluable assistance in arranging the seismic survey in Loch Linnhe, for lucid and incisive advice on several occasions and I acknowledge considerable assistance from him in the interpretation of the seismic results.

Mike, Nick S. Anna, Nick H., Claire, Alastair, Norman and Al have all provided invaluable field assistance and excellent company which helped restore my sanity during long periods of camping alone in Western Lochaber. Bogs were expertly cored by Richard Tipping, Bob and Anthony. I also thank Gavin Aitcheson in the air photo unit for cheerful assistance during months of visits, and Mikko Punkari for digitising the seismic results. This thesis has benefited considerably from discussions with, and assistance from, Mike, Gordon, Al, Alastair, Alun, Nick S., Andy K., Peter Thorp and Charles.

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Chapter 1 - Introduction

1.1 Aim

The aim of this thesis is to use field evidence to reconstruct the extent of the Loch Lomond Stadial ice cap in Western Lochaber to infer controls on former ice dynamics and to shed light on Quaternary landscape evolution.

1.2 Rationale

Reconstruction of the Loch Lomond Stadial ice mass in Scotland is important for a variety of reasons.

Lochaber is part of Western Scotland, in North West Europe (Fig 1.1). Scotland is an ideal location for studying the terrestrial response to climatic changes. This is because both theoretical and empirical evidence shows that the British Isles are a particularly sensitive region in terms of glacier responses to climatic cooling. This sensitivity is due to Britain's maritime climate which is suitable for rapid glacial growth and decay, and as Scotland is currently close to the glaciation threshold (Payne and Sugden 1990a, Kerr 1992). Scotland is thus likely to have a relatively high magnitude glacial response to any climatic trigger. In addition, the limited size of the British Isles mean that only a relatively small ice sheet can develop (in comparison to the Laurentian or Fennoscandian ice sheets, for example), and small ice sheets respond relatively rapidly to climatic changes. Scotland's environment can thus potentially respond to high frequency, low magnitude climatic events, allowing high resolution models of the mechanisms involved to be tested.

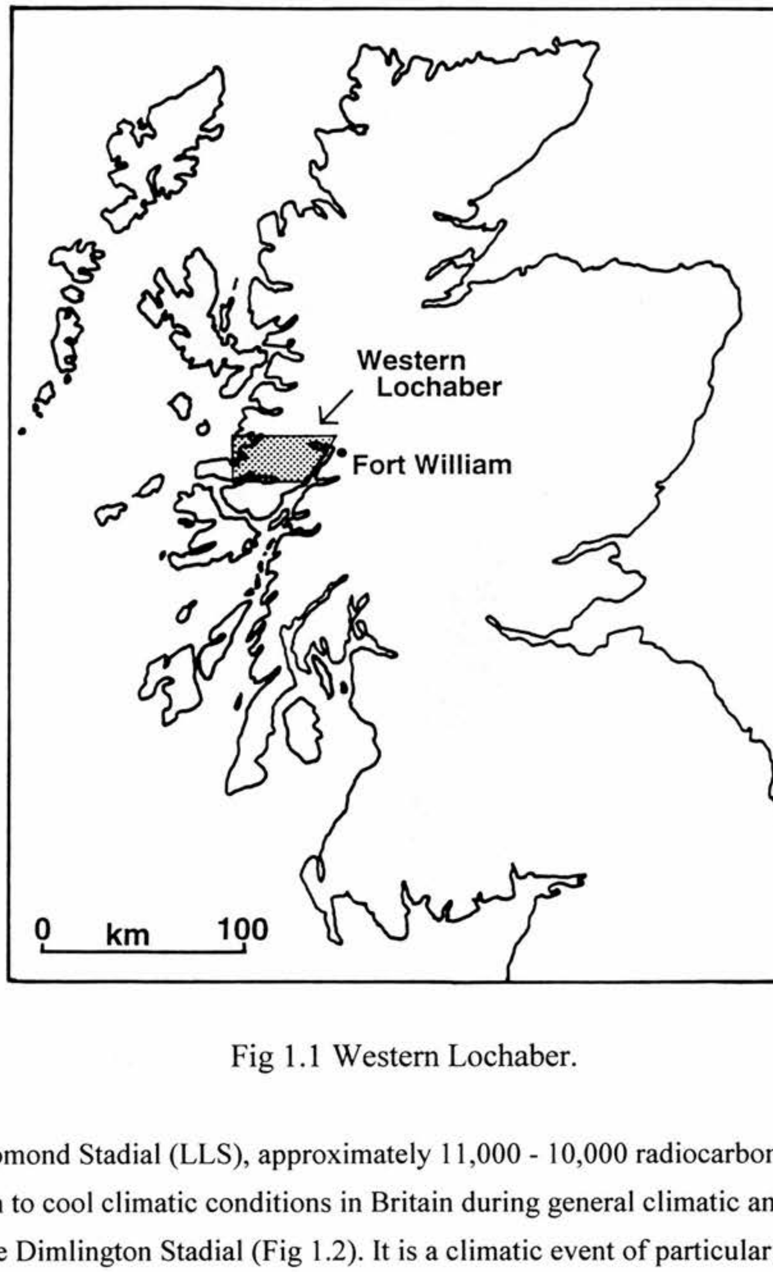

Fig 1.1 Western Lochaber.

The Loch Lomond Stadial (LLS), approximately 11,000 - 10,000 radiocarbon years BP, was a brief return to cool climatic conditions in Britain during general climatic amelioration following the Dimlington Stadial (Fig 1.2). It is a climatic event of particular interest for several reasons. Firstly, the LLS glaciation was the most recent period when there was a markedly cooler climatic regime to that of today. Evidence of contemporary environmental conditions during this cooling event will thus be the best preserved. This allows relatively well constrained reconstructions both of palaeoclimates and of palaeo-environments. Secondly, the Stadial offers the unusual potential to examine a complete cycle of a glacial response to a short, abrupt episode of climatic cooling. This is in contrast to studies of contemporary glacial fluctuations and climatic change, for example, where changes are

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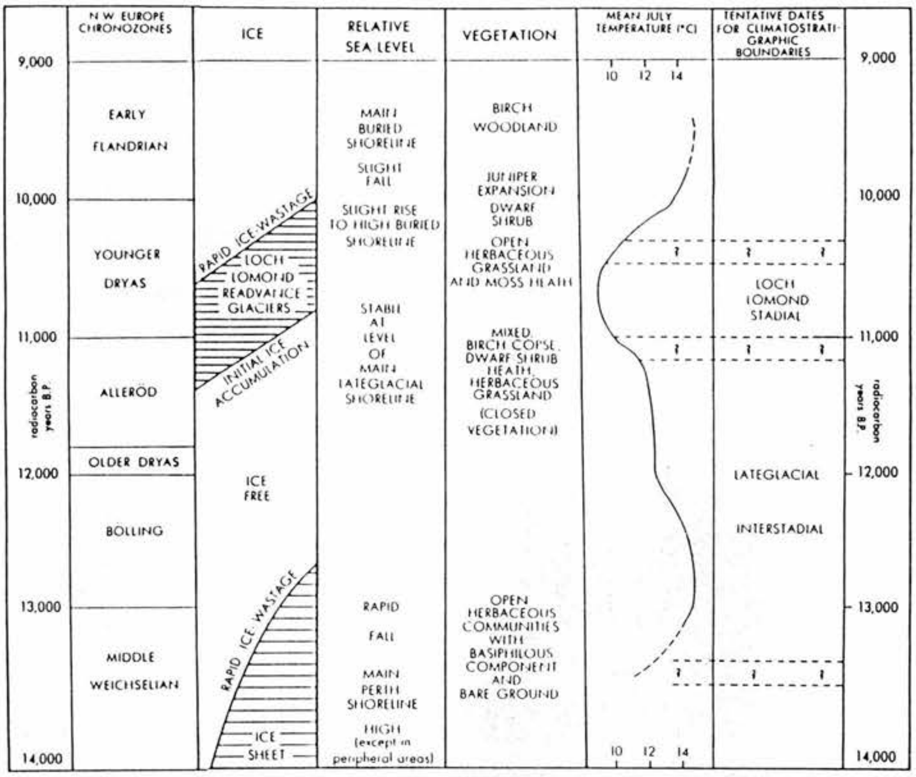

Fig 1.2 Lateglacial temperature, ice volume and environmental changes in Western Scotland, from Gray and Lowe (1977).

? qualify

affecting pre-existing ice masses with various lag times. Thirdly, since the LLS probably represents a truncated glaciation, the processes of ice growth can be investigated; evidence for glacial advance is normally removed during a full ice sheet phase (Kerr 1992). Fourthly, the LLS offers the potential to investigate the processes involved in glacial-interglacial transitions which are not yet fully understood. In particular, the reasons for the brief return to cool, yet fluctuating (Taylor et al. 1993), climatic conditions during the Younger Dryas in NW Europe (the LLS in Britain), and possibly globally (Broecker 1994, Denton and Hendy 1994) are not established. In NW Europe, this event is associated with a renewed southwards movement of the North Atlantic Polar Front (Ruddiman and MacIntyre 1981), possibly induced by a large influx of icebergs melting and releasing freshwater into the North Atlantic (Broecker 1994). It is unknown whether this movement of the polar front is a cause or a response to global cooling, although North Atlantic circulation is thought to be crucial to global climatic change (Sutherland and Gordon 1993). Accurate knowledge of LLS climatic and environmental conditions in Scotland, where climatic conditions are largely controlled by frontal systems passing over the North Atlantic, are thus important in testing theories of the mechanisms involved.

The high magnitude, discrete nature of the LLS glaciation in Western Scotland offers an ideal case study with which to investigate lithospheric/cryogenic/oceanic/atmospheric interactions. Modelling experiments studying the glaciological response to changes in climatic variables and the effects of topography have used the LLS ice sheet as a case study with which to constrain the model parameters (Payne and Sugden 1990b, Kerr 1992). Lambeck (1991, 1993a,b, 1995) has used the extent of late Devensian and LLS ice over the British Isles as input to a high resolution model which infers lithospheric and mantle characteristics from empirical evidence for sea level changes due to ice and meltwater loading and unloading. An accurate, empirically based, three-dimensional reconstruction of the former ice mass is essential for testing and calibration of these models. There have been few empirically based attempts to reconstruct the three-dimensional LLS ice surface profile over many parts of Western Scotland, including Western Lochaber, so that these models have used insecure reconstructions (eg. Sissons 1974b, 1976, 1981). With better constraints, such models should allow better understanding of the factors controlling previous rates and processes of environmental change, and hence an improved framework for assessing current changes and the impact of human activities on the environment.

Reconstruction of LLS ice in Western Lochaber is of interest for a further reason. This area is deeply dissected by sea lochs and the LLS ice cap would have been drained almost entirely by tidewater glaciers. Both empirical evidence from Alaska and elsewhere and theoretical considerations suggest that the behaviour of glaciers with calving termini is different to that of those with terrestrial termini. In particular, at a scale of 500 - 1,000yrs, their glacial response to climatic changes tends to be filtered through several non-climatic controls on ice dynamics, so that these glaciers undergo cycles of advance and retreat which are controlled by glaciodynamic mechanisms (Mann 1986, Mayo 1988). The most important of these non-climatic controls are sea-level changes, trough geometry and sediment flux to the terminus, and these may have a stronger influence on ice extent and dynamics than climate, at time scales of 10's-100's of years (Powell 1984, Mercer 1961). As the LLS glaciers were probably in an extended position for only a few hundred years (Payne 1988), this area offers an ideal example to study how the responses of tidewater glaciers to climate are affected by these controls.

In the past it has been assumed that the areal extent of LLS ice was a direct product of palaeoglacier mass balance parameters such as climate, aspect and snow input by avalanching. This assumption forms the basis of attempts to calculate former ELAs of reconstructed glaciers, and to make palaeoclimatic inferences on the basis of these ELAs and the distribution of ice with aspect (e.g. Sissons 1974a,b, Sissons and Sutherland 1976, Ballantyne 1989). Recent work suggests that views of a simple relationship between climate and ice extent are incorrect; topography (Kerr 1993) and tidewater glacier dynamics, in particular, may exert strong additional controls on glacier dimensions. Reconstruction of the LLS ice cap and dynamics in Western Lochaber can potentially assess the ways in which reconstructions of ice extent are used to infer palaeoclimates.

1.3 Physical geography of Western Lochaber

Topography

Western Lochaber consists of mountainous topography dissected by numerous troughs (Fig 1.3). The main mountain summits are mostly around 750m altitude, although those in the north reach 950m. Several of the troughs in Ardgour are breaches which cut across the main watershed. In several instances the watershed may be at corroms (eg. Glen Tarbert), where the direction of water flow from a hanging valley at the watershed is controlled by the detailed pattern of alluvial fan aggradation. Many of the troughs contain sea or freshwater lochs. Basins in Lochs Linnhe and Shiel are more than 100m deep and Lochs Eil and Sunart attain depths of 70-90 m, whereas Lochs Ailort and Moidart are less than 30m deep. Loch Linnhe is the largest and deepest loch and follows the fault shatter belt of the Great Glen (Bailey and Maufe 19~~60~~ 16). In the extreme west and south of the area there are plateaux at altitudes of 3-400m. These have irregular, low relief surfaces with numerous lochains.

Geology

The bedrock of Western Lochaber comprises Moinian schists and gneisses, with numerous local igneous intrusions of various dimensions, ages and lithologies (Fig 1.4). The Moine supergroup consists of metasedimentary rocks, and has been subdivided into the Morar, Glenfinnan and Loch Eil groups, which are represented in Western Lochaber from west to east respectively. These groups are now known to be in stratigraphic succession, the Loch Eil division being the youngest (Harris 1991). A belt of granitic gneiss runs through Ardgour, and is of migmatitic origin (Dalziel 1966), contemporary with the tectonic and metamorphic events which affected the country rocks into which it was emplaced (Harris 1991).

The main significance of bedrock geology in this thesis is in the resistances of different lithologies to weathering, periglacial action and glacial erosion. The Moinian schists and gneisses range from massive gneisses, quartzites and siliceous psammites to micaceous

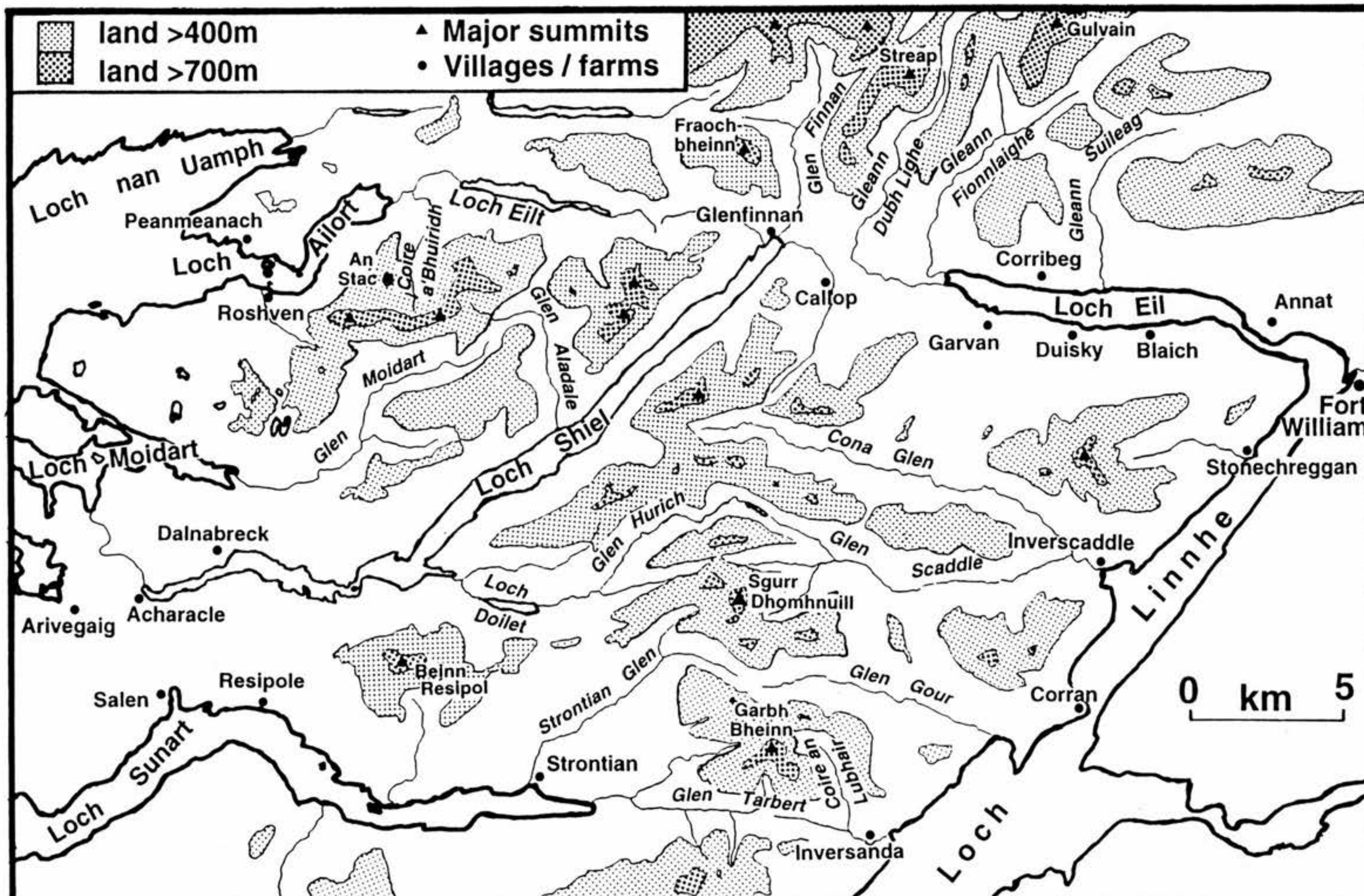

Fig 1.3 Topography of Western Lochaber, showing locations of places mentioned in the thesis.

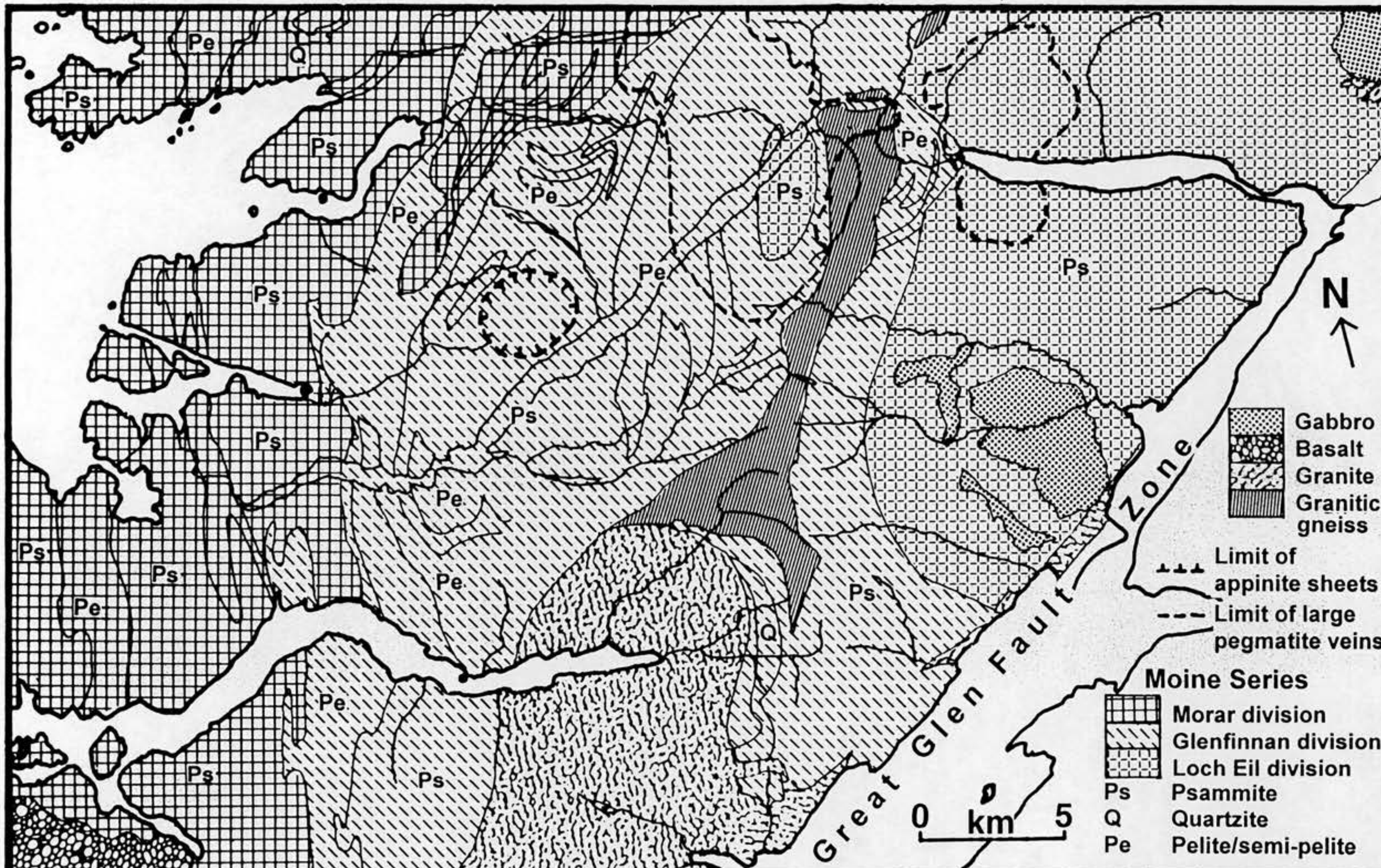

Fig 1.4 Geology of Western Lochaber.

semipelites and pelites. The micaceous pelites and semi-pelites show a well developed fissility, manifested in the field as closely spaced lines of weakness which have been exploited by frost action to form joints and cracks in the bedrock surface. Clasts formed by macrogelivation are normally platy and flaggy shaped. Micaceous rocks also commonly have rounded edges suggesting they have been subject to microgelivation. The psammites, gneisses and siliceous semi-pelites are less well jointed in the field, and the clastic products of frost shattering tend to be block shaped. Some quartzites are particularly resistant to edge rounding, and retain angular edges.

The largest igneous intrusion is that of Strontian granite. This underlies a broad topographic low, is very coarse grained and is weathered to form regolith up to 30cm thick. There is another large intrusive complex at Inverscaddle. Western Lochaber is crossed by numerous igneous vein complexes and minor intrusions of a variety of lithologies and ages. In the field many of these lithologies are frost shattered into small blocks, for example the fine grained acid felsites, and the coastal granite along Loch Linnhe.

There are also numerous faults in Ardgour, some trending NE - SW which are thought to be associated with fault activity in the Great Glen Fault, and others trending E - W, NW - SE, and N - S (MacGregor 1967).

1.4 Landscape evolution in Lochaber

Pre-Quaternary

As ice sheets remove most evidence of pre-existing environments, evidence for pre-Late Quaternary landscapes and geomorphic processes is mostly inferred from the sedimentary evidence in offshore basins. This evidence is fragmentary and many interpretations remain speculative. It is likely that the first Quaternary ice sheets built up on a landscape which was largely similar to that of today in terms of the location of valleys and mountains. Indeed evidence suggests that there has been a continuity of relief development over the last 400Ma (Hall 1991). The location of valleys is probably strongly influenced by faults and lines of geological weakness (MacGregor 1967, C. Le Coeur unpublished data, 1994), although early geological work sometimes inferred the presence of faults from topographic features rather than lithological characteristics, thus inducing problems of circular reasoning. The plateaux at 300-400m in the west and south of Western Lochaber are former erosion surfaces and may be diachronous (Hall 1991).

Warm, humid conditions during the Tertiary resulted in deep chemical weathering, the depth of regolith created being strongly dependent on lithostructural controls (Hall 1991). Strontian granite, for example, is likely to have been particularly susceptible to Tertiary weathering. Tertiary volcanic activity and associated tectonic uplift in the Inner Hebrides probably resulted in the Lochaber palaeo-coastline being further west than that of today, as the igneous centres of Mull and Rhum may have been peninsulas rather than islands prior to Quaternary glacial erosion (Hall 1991). The pre-glacial drainage pattern in Western Lochaber is unclear (Hall 1991). Some reports have suggested the watershed was west of that of today, so that much of Scotland was characterised by eastwards flowing drainage patterns (e.g. Bailey and Maufe 1960). The low cols in Ardgour have been interpreted as relicts of this eastwards flowing drainage system. In contrast, Sissons (1967) suggested that the low cols are partly remnants of rivers flowing westwards across the Great Glen fault from mountains to the east. Others, however, have argued that the low cols in Ardgour are at least partly the product of Quaternary glacial erosion, and thus may not necessarily be indicative of pre-glacial river valleys (e.g. Linton 1949). All these ideas imply that the Linnhe trough may not have been excavated to its present depths before the Quaternary. At the end of the Tertiary, then, the landscape of Western Lochaber probably featured mountains and river valleys in the same locations as those of today, and a thick weathering mantle over the land surface.

Early - mid Quaternary

During the Quaternary there is evidence for 3-5 ice sheet glaciations in Scotland, and there were probably several smaller highland glaciations (Sutherland and Gordon 1993). The glacial features widespread in the highlands today are thus the product of repeated episodes of glacial erosion. It is likely that the first ice sheet caused the most extensive landscape modification, as there would have been large amounts of easily eroded Tertiary weathered regolith covering the land surface (Sutherland and Gordon 1993). In addition, the landforms would have deviated most from glacier equilibrium forms and would have undergone rapid initial transformation (Sugden and John 1976). These ice sheet and highland glaciations sculpted the main overdeepened troughs, hanging valleys, arêtes and corries present in the landscape today, and deposited thick sequences of glacigenic deposits offshore. Quaternary glacial erosion in dissected mountainous areas like Skye is estimated as ~ 120m of valley basal lowering with 100-250m of back wall retreat depending on rock jointing (Le Coeur unpublished 1994); and valley deepening is up to ~ 200m at Loch Leven (Borthwick, pers. comm.). More limited landscape alteration is suggested in less dissected areas, such as low plateaux, by surface lowering of up to 50m due to removal of the pre-glacial weathering regolith (Hall 1991).

Devensian ice sheet

The last (Devensian) ice sheet is commonly believed to have reached a maximum extent during the Dimlington Stadial (26,000 - 13,000 BP). The extent and dynamics of this ice sheet are the subject of some controversy (Boulton et al. 1977, 1985), but the ice sheet probably reached its maximum extent asynchronously at different margins between 25-18,000 BP. It extended into the North Sea, to Norfolk, and coalesced with an Irish ice sheet. Off Western Scotland, it coalesced with an ice sheet based over the Outer Hebrides. Devensian ice overrode the highest mountain summits over mainland Scotland south of Inverness, probably attaining a maximum thickness of 1300 - 1700m (Lambeck 1993b, Glasser 1992).

In Western Lochaber, field evidence suggests that Devensian ice flow in Moidart and Sunart was westwards (Peacock 1970, BGS 1:50,000 drift sheets 52 and 61), south westwards in Western Ardgour (Bailey and Maufe 1960, Thorp 1987) and south eastwards in Eastern Ardgour (Bailey and Maufe 1960). In several instances Devensian ice flow direction indicators suggest ice movement over ridges and summits, suggesting that ice flow at the maximum may have been to some extent independent of local topography.

Deglaciation occurred between 18,000 to 13,000 BP, and during this time the summits and slopes would have been progressively exposed to periglacial, weathering, mass wasting and fluvial processes.

Lateglacial Interstadial

Scotland may have been entirely ice free during the Lateglacial Interstadial from 12,500 - 11,000 BP (Fig 1.2), although it has never been proven that ice did not survive this period in the high corries. Palaeo-environmental indicators suggest that temperatures in Scotland rose rapidly at the start of the Interstadial to within a few degrees of those at present (Coope 1977), as North Atlantic Drift waters moved northwards to affect British coasts. Vegetation became re-established, and weathering, mass wasting, periglacial and fluvial processes similar in magnitude and intensity to those of today are inferred.

Loch Lomond Stadial

The Loch Lomond Stadial was a brief return to cold arctic conditions from ~11,100 - 10,000 BP, associated with the re-establishment of polar waters around Britain. This resulted in the regrowth of ice masses in topographically and climatically suitable locations. Small corrie and valley glaciers readvanced in the English Lake District, the southern Uplands, parts of the Eastern Cairngorms, some Inner Hebridean islands and the far north of Scotland. The main body of ice was an ice cap in Western Scotland (Fig 1.5).

This separate glacier readvance has been recognised from the field evidence for over 100 years, and in some areas the ice limits have been reconstructed in detail. In many areas this is facilitated by prominent end moraines (Sissons 1979a), glacial trimlines (Thorp 1981), and outwash features (Thorp 1986). In many places smaller hummocky moraines abound within the limits and have been attributed to LLS ice activity. Recent work suggests that, in probably resulted in the Lochaber palaeo-coastline being further west than that of today, as the igneous centres of Mull and Rhum may have been peninsulas rather than islands prior to Quaternary glacial erosion (Hall 1991). The pre-glacial drainage pattern in Western Lochaber is unclear (Hall 1991). Some reports have suggested the watershed was west of that of today, so that much of Scotland was characterised by eastwards flowing drainage patterns (e.g. Bailey and Maufe 1960). The low cols in Ardgour have been interpreted as relicts of this eastwards flowing drainage system. In contrast, Sissons (1967) suggested that the low cols are partly remnants of rivers flowing westwards across the Great Glen fault from mountains to the east. Others, however, have argued that the low cols in Ardgour are at least partly the product of Quaternary glacial erosion, and thus may not necessarily be indicative of pre-glacial river valleys (e.g. Linton 1949). All these ideas imply that the Linnhe trough may not have been excavated to its present depths before the Quaternary. At the end of the Tertiary, then, the landscape of Western Lochaber probably featured mountains and river valleys in the same locations as those of today, and a thick weathering mantle over the land surface. D

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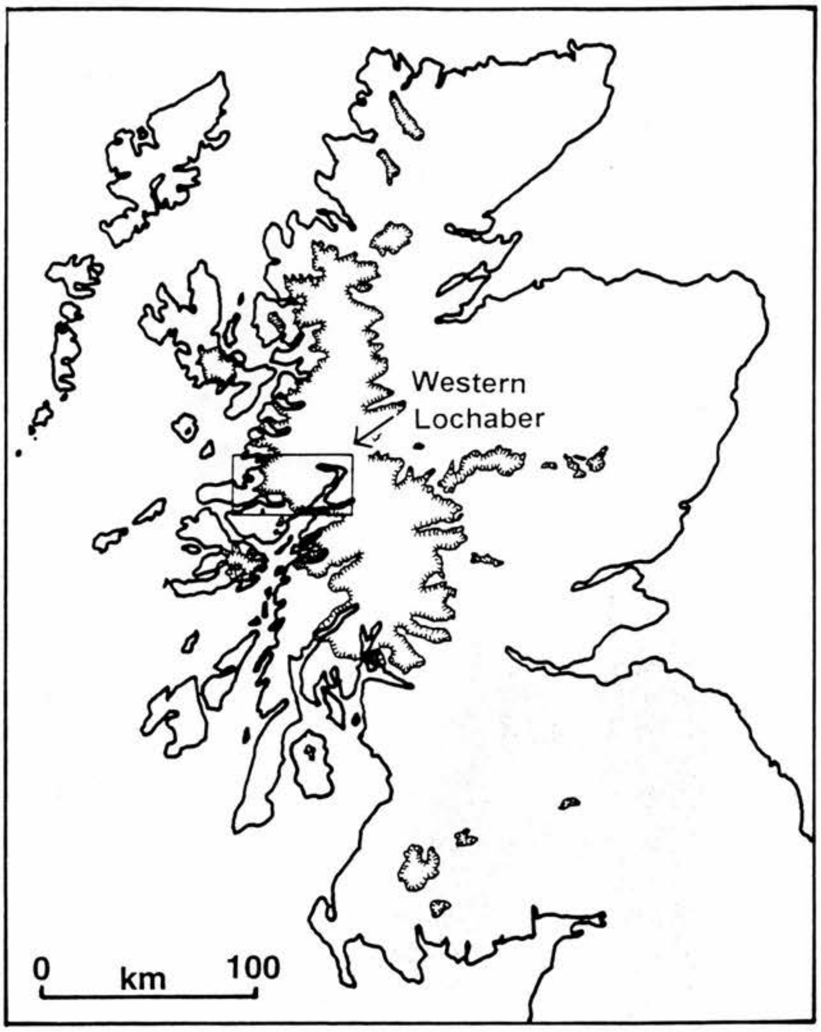


Fig 1.5 Loch Lomond Stadial ice limits in Scotland.
Sources: Ballantyne 1989, Bennett 1991, Gray 1975, Sissons 1967,
Sutherland 1984, Thorp 1986.

some locations, the order present in the spatial arrangement of these features can be used to infer former ice marginal positions and the patterns of LLS glacial retreat (Benn 1992, Bennett and Boulton 1993a,b), although the exact timing of deglaciation remains uncertain due to dating problems. In much of the western half of the former ice cap, however, clear terminal moraines are absent; the area was referred to by Charlesworth (1955) as the 'moraineless west'. Partly due to the difficulty of establishing former glacial limits in this situation, there have been few detailed field studies of LLS glacial features and few field-based attempts to reconstruct the LLS ice limits in many western coastal areas, including Western Lochaber. Most of the detailed mapping and reconstruction of LLS ice limits completed to date was instigated by Sissons and his co-workers 15 - 30 years ago. Recent

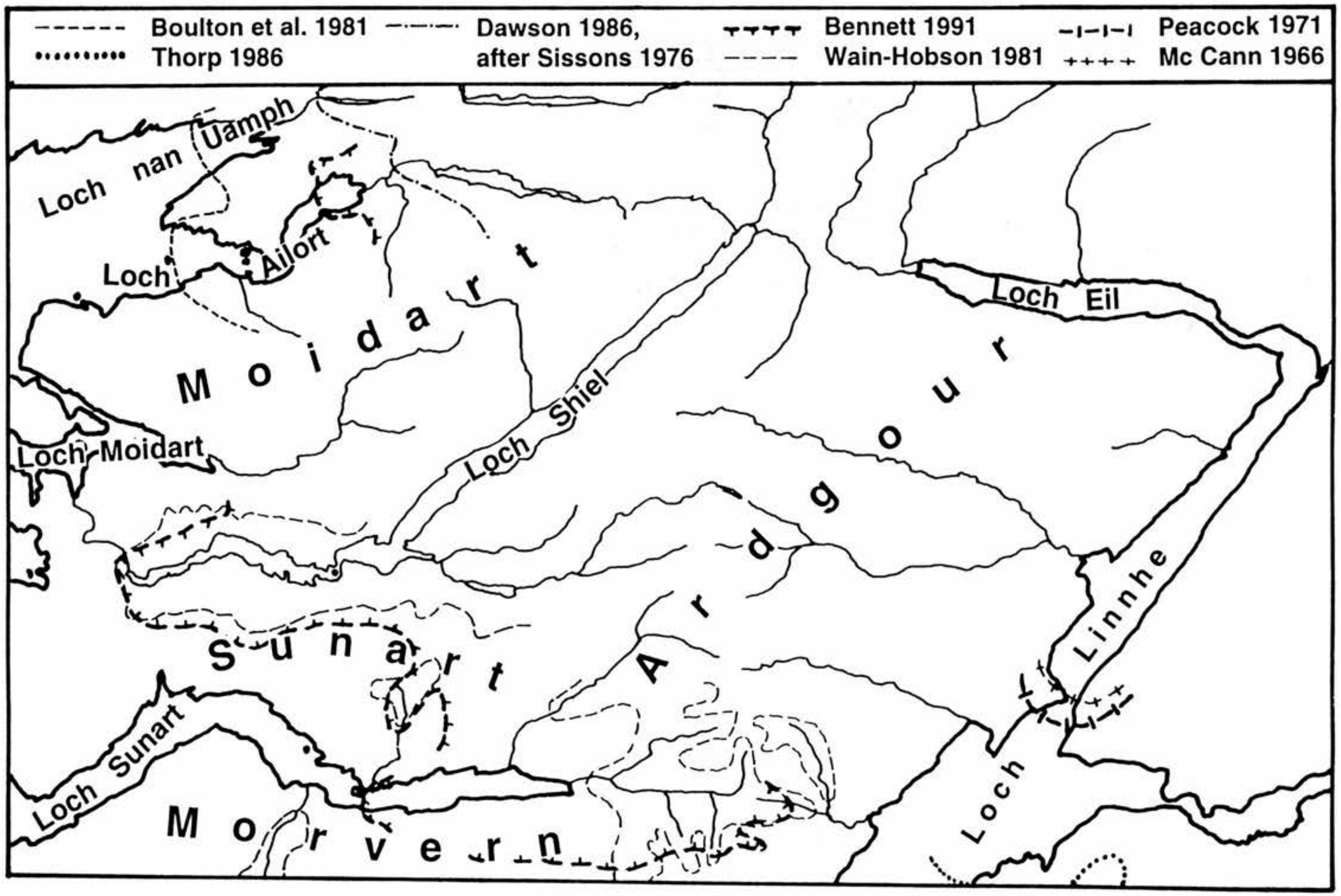


Fig 1.6 Different interpretations of Loch Lomond Stadial ice limits in Western Lochaber.

Sources: Bennett 1991, Boulton et al. 1981, Dawson 1986, McCann 1966, Peacock 1971, Thorp 1986.

studies using advances and refinements in glaciological, geomorphological and sedimentological theory have sometimes led to revision of the previously reconstructed ice limits, mostly increasing the reconstructed ice extent (Ballantyne 1989, Bennett 1991, Benn and Evans 1993, Thorp 1986).

Some of the previously proposed LLS ice limits in Western Lochaber are shown in Fig. 1.6. These limits have not been the subject of detailed field studies in the past; some terminal and retreat features have been identified, but there have been no previous attempts to investigate ice thicknesses in the interior of the area (Fig 1.6). The variable ideas of ice extent in this area are shown in recently published maps of Scottish LLS ice limits (e.g. Gray and Coxon 1991, Firth et al. 1993, Bennett and Boulton 1993a). Ice limits in Sunart, Southern Ardgour and Morvern were proposed by Wain-Hobson (1981). Several authors have suggested that the Shiel glacier terminated at an ice contact slope near Acharacle (Peacock 1970, McCann 1966, Wain-Hobson 1981, Charlesworth 1955, Bennett 1991). Early workers suggested that the LLS Linnhe glacier terminated around the Corran outwash fan (Peacock 1970, McCann 1966, Wain-Hobson 1981). This was later re-interpreted as the location of a stillstand during glacier retreat by Thorp (1986), who discovered further LLS glacial outwash deposits a few kilometres down loch in the course of mapping the LLS ice limits of Eastern Lochaber. Boulton et al. (1981) used seismic evidence in Loch Ailort to suggest that the LLS Ailort glacier extended to the mouth of the Loch. In contrast, Dawson (1988, 1994a) supports Sissons' (1976) tentative limits which show LLS ice terminating east of Lochs Ailort and Moidart. Dawson's support for these limits is based on the distribution of raised marine features and an interpretation about the age and genetic origin of one depositional feature which is at odds with that presented later in this thesis. Charlesworth (1955), Sissons (1976, 1983) and Bennett (1991) have all suggested outer ice limits in the area, but these were largely based on extrapolation from clearer ice limits elsewhere, rather than empirical evidence in the area. In addition, Bennett (1991, Bennett and Boulton 1993a, 1993b) have reconstructed the LLS deglaciation pattern throughout the Western Highlands based on air photograph interpretations of assemblages of ice marginal features present within hummocky moraine (Fig 1.7). This work was not subject to detailed field checking in Western Lochaber.

The severe climate of the Stadial resulted in harsh environmental conditions outside the ice limits. Non-glacial geomorphological processes were particularly active, leading to rockfalls, the formation of protalus ramparts, talus and rock glaciers in the mountains (Sutherland and Gordon 1993). There was widespread permafrost and periglacial activity at higher altitudes which resulted in mountain-top detritus, solifluction lobes, sorted circles, stripes and intensely frost shattered rock, with the particular suite of periglacial features formed being dependent on local lithological and regolith characteristics (Ballantyne 1984). There is also evidence of enhanced fluvial activity and slope action (Sutherland and Gordon 1993). Enclosed depositional basins received mainly minerogenic inwash during the Stadial. This inwash usually contains little plant pollen, and that present is derived from species typical of tundra and unstable ground (Lowe and Walker 1984).

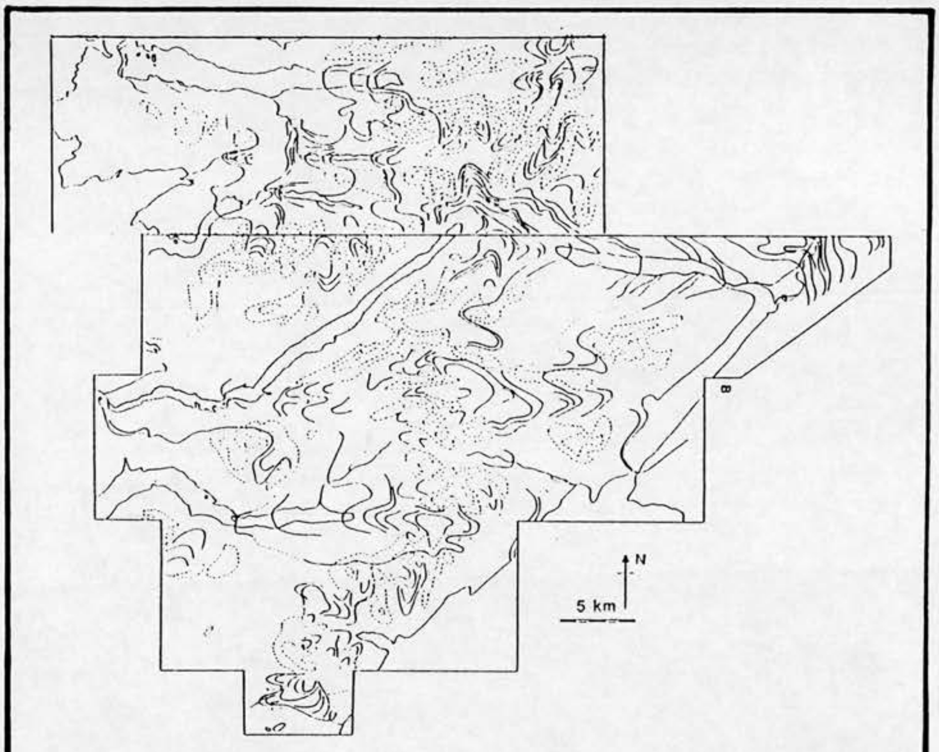


Fig 1.7 Loch Lomond Stadial deglaciation in Western Lochaber, from Bennett 1991.

Holocene

A variety of palaeotemperature indicators point to rapid warming at the end of the LLS. This scenario was originally proposed by Coope (1977) on the basis of beetle evidence, and is now supported by the evidence of sudden changes of up to 7^{o}C in 50 years or less indicated in Greenland ice cores (Dansgaard et al. 1989). As the climate ameliorated vegetation and soils became established and this led to a general diminution in the intensity of geomorphic activity. Tectonic movements associated with ice unloading, and the large amounts of unstable glacial debris on slopes, however, meant that in the early Holocene, landslips, slope failures, and the reworking of debris by streams and debris flows into river terraces and debris cones was probably rapid (Holmes 1984, Ringrose 1987, Sutherland and Gordon 1991). Limited periglacial activity on mountain summits at present produces small scale forms, dependent on local regolith (Ballantyne 1984, 1987).

Sea-level changes

As global and local ice sheets have waxed and waned, and the Earth's lithosphere adjusted to the spatial redistribution of mass caused by these changes in ice volume and sea water, a complex pattern of spatial and temporal sea-level changes around the British Isles coastline has evolved.

In Western Scotland, relative sea-levels have been falling since the retreat of Devensian ice, except for a transgression in the early-mid Holocene. Raised marine features are thus present around the coastline, different altitudes being associated with different ages. Attempts to establish sea-level curves in the west have been hampered by the poorly developed shoreline features, the difficulty of relating different raised marine features to particular sea-levels, problems finding closely spaced evidence (Lambeck 1993a), and the lack of dates on raised beach and erosional features. In particular, controversy has long reigned over the age of the Main Rock Platform in Western Scotland (e.g. Peacock 1975, Sissons 1975, Gray 1989, Dawson 1989). Many have interpreted it as a result of rapid rates of coastal erosion during the severe environmental conditions of the LLS. Yet there is a growing body of evidence that the feature was originally formed well before the LLS (Gray and Ivanovitch 1988) and retrimmed to some extent during this period (Peacock et al. 1978, Thorp 1984). Until the age and origin of this feature is more securely pinned down, it is probably unwise to use it as evidence for LLS sea-levels.

Shennan et al. (1993, 1994a, Shennan 1994) have recently used detailed investigations of radiocarbon-dated lithological, palynological and diatom stratigraphies of isolation basins close to the West coast of Lochaber to reconstruct a well constrained lateglacial sea-level curve for this area (Fig 1.8). This shows that sea-levels were sufficiently high during the LLS for Loch Shiel to have been a sea loch (present altitude 4m O.D.). Prior to 12,000 BP sea-levels probably fell rapidly from ~40m O.D. during Devensian deglaciation (Peacock 1970, Lambeck 1991) to a minimum of 4m O.D. during the late LLS and early Holocene. During the Postglacial Marine Transgression (PGMT) sea-levels rose again to ~7m O.D. , and have been falling subsequently. Lateglacial sea-level changes in Loch Linnhe and the Firth of Lorne are less well established; there was probably a similar pattern of change to that on the west coast, but at different altitudes. Sea-level in the Oban region fell from ~34m O.D. during Devensian deglaciation to ≤ 10-12m O.D. during the LLS, and then re-attained this altitude during the Postglacial marine transgression before continued fall during the rest of the Holocene (Benn and Evans 1993, Gray 1975, 1994, Peacock 1970, Lambeck 1993b).

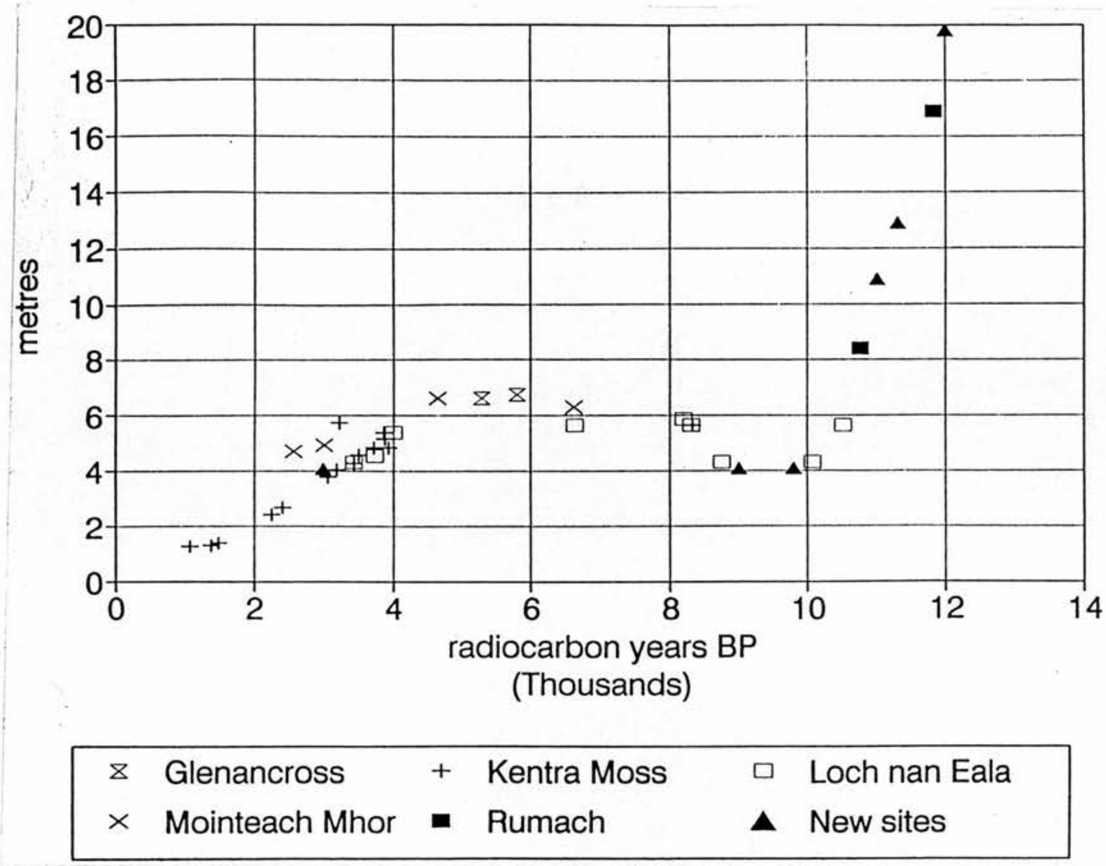


Fig 1.8 Lateglacial and Holocene mean sea-level changes at sites near Kentra and Arisaig, from Shennan 1994. Different symbols indicate different sites.

1.5 Thesis structure and organisation

Here the areal extent and dynamics of the LLS ice cap in Western Lochaber are reconstructed using a variety of different forms of evidence. Air photograph and field geomorphological mapping is used to establish terrestrial glacial limits in the lower reaches of the former glaciers, to examine spatial variations in the types of glacial evidence present and to infer the locations of stillstands during glacial retreat (Chapter 2). Chapter 3 presents the results of 111 trimline mapping traverses up hillslopes in Western Lochaber which are used to suggest ice surface altitudes in the upper and middle reaches of the former glaciers. In Chapter 4 the results of a seismic survey in Loch Linnhe are presented. This provides essential evidence with which to corroborate terrestrial and marine glacial limits. Chapter 5 synthesises all three forms of evidence and presents the reconstructed LLS at the ice maximum. Evidence supporting the age of the ice cap is discussed and the deglaciation pattern is reconstructed. Chapter 6 assesses the implications of the reconstruction, including the role of tidewater glacier dynamics and palaeo-ELAs. The main conclusions and wider significance of the thesis are summarised in Chapter 7.

Chapter 2 - Landform evidence

2.1 Aim

This chapter describes and interprets the terrestrial geomorphological evidence for glacial, periglacial and raised marine action in Western Lochaber.

2.2 Methods

2.2.1 Techniques

Western Lochaber was first examined on 1:10,000 1940s and 1:24,000 1988 black and white vertical air photographs. The 1:10,000 series have the advantage of pre-dating much of the afforestation in the area, and there is also often more than one run covering any one area, which compensates for variations in shadow and lighting conditions. The 1:24,000 series are very good quality, high resolution images, taken at times of the day and year to minimise loss of coverage due to shadows. The geomorphological features mapped from the air photographs were transferred on to 1:25,000 base maps, with additional 1:10,000, or 1:5,000 base maps of small areas where there are interesting features requiring a higher level of detail. These maps were taken into the field for intensive field checking of areas that have not been afforested or developed, and were modified where necessary. Small features such as striations and periglacial forms were mapped directly in the field. Profiles were surveyed over significant landforms using an Abney level and tape, and heights and altitudes were obtained either by Abney level, or using an Electronic Distance Measurer, according to the accuracy required. Where altitudes above a datum were surveyed with an Abney level, the datum used was Mean High Water Springs (MHWS), which was taken as the highest large line of seaweed on a beach, or Mean Sea Level (MSL), which was inferred from the distribution of seaweeds and lichens.

Sedimentary evidence was collected from depositional landforms by examining exposed sedimentary sections. These are found alongside forestry and estate tracks, roads, paths, quarries and sometimes in stream and river banks. All the roads and tracks in the field area were examined for bank sections. Notes were taken of the sedimentary characteristics of all visible sections, and a basic sedimentary log and two dimensional sketch of important ones were made in the field, with the aid of a polaroid camera. In some instances measurements of pebble fabrics, clast roundness, shape or lithology were recorded. Where clast shape and Why? (Appendix 1A?) roundness were measured, samples of 50 clasts with a axes between 3 and 12 cm were used A1 from within an area of $1m^2$, and roundness assessed using the criteria given by Benn and Ballantyne (1994). Lithological analyses also used samples of 50 clasts. Why?

Several types of raised marine evidence were collected. The altitudes of raised beach bars, the backs of raised beaches without clear bars and marine erosional notches in outwash or till give evidence for former sea-levels. In other situations the lowest altitude at which subaerial sediments are present can give a maximum altitude for sea-level at the time of deposition, and maximum altitudes of marine sediments can similarly provide a minimum value. These latter forms of evidence may be less reliable than the former, as it is difficult to definitively distinguish between submarine and subaerial sediments deposited in some environments.

As different types of marine geomorphological and sedimentological features are formed at different altitudes with respect to the contemporary mean palaeosea-level (MSL), the measured altitudes of each palaeosea-level indicator were standardised to the implied MSL, in metres above O.D. Newlyn (Lambeck 1993b, Shennan et al. 1994a, Shennan 1994, van de Plassche 1986). It is assumed here that storm beach bars form at MHWS + 0.5 - 2.5m, depending on the size of the bar and the exposure of the coastline. Most other forms of evidence are assumed to be related to MHWS (Table A.3.1). Conversion to the Newlyn datum from an altitude measured with reference to current MHWS or MSL was made using tidal information obtained mainly from Admiralty charts, summarised in Table A.3.2 in Appendix 3. Errors in inferred palaeosea-levels are discussed in Section 2.4.

2.2.2 Criteria for identifying glacial features

Glacial features were identified by comparison with previously published descriptions of the characteristics of glacial landforms and sediments.

Glacial moraines

There has been much debate over the genesis of Scottish hummocky moraine over the last decade. However, the most recent work on moraines by Benn (1992) shows that, on Skye, the landform is polygenetic, with three sediment-landform assemblages recognised (Table 2.1).

Table 2.1 Hummocky moraine classification, after Benn (1992).

Moraine	Recessional	Chaotic	Drumlins and fluted
Orientation	transverse ridges in belts of sub-parallel ridge suites	random	longitudinal bedforms
Geomorphology	ridges and chains of hummocks	varies - conical hummocks, composite hummocks, flat topped mounds, rim ridges	ridges 1-10m high, and up to 400m long
Sediments	variable, often diamictons showing evidence of fluvioglacial and debris flow activity. May have trust faults and folds	debris flow and mass movement deposits interbedded with water sorted material. many signs of sediment reworking. May have normal faults and slumps.	compact, sheared lodgement till with edge rounded and faceted clasts
Genesis	debris accumulation and glacitectonics at an active ice margin	deposition in contact with stagnant or slowly moving ice (may only be local)	subglacial bedforms

The first are recessional moraines comprising transverse push and dump ridges, and recording the ice front positions of an actively retreating ice margin. Bennett and Boulton (1993a, 1993b) consider that most hummocky moraine fragments in the Scottish Highlands, when viewed on aerial photographs, can be linked together to form chains of mounds and ridges, which can be interpreted in this manner as evidence of former ice front positions of an actively retreating ice mass. Secondly, Benn recognised moraines with a chaotic spatial arrangement, resulting from uncontrolled ice-contact deposition or stagnation. These can occur in close proximity to recessional moraines, and Eyles (1983) noted that apparently chaotic moraines can form in front of active glaciers; consideration of the wider context in which these moraines are located is essential before making inferences about former ice dynamics, as the presence of these moraines can only give evidence of local ice stagnation. Until recently, it was widely believed that most Scottish hummocky moraine was chaotically arranged, and formed by deposition from stagnant, inactive ice (eg. Sissons 1967, 1976, Sugden 1970). Finally, there are longitudinal drumlins and fluted moraines which are formed subglacially, as described in Torridon by Hodgeson (1982). These three categories of hummocky moraine are distinguished on the basis of morphology, spatial arrangement, and internal sedimentary characteristics as shown in Table 2.1.

Glacial sediments

Numerous recent investigations have focused on the wide range of processes involved in till formation, and the equally wide range of end products (e.g. Hart and Boulton 1991, Krzyszkowski 1994). This study uses a simple three-fold classification of tills composed mainly of supraglacially transported material released by ice marginal meltout, tills composed mainly of subglacially transported material deposited beneath the former glaciers, and tills which have undergone post-depositional resedimentation by sediment gravity flows. These tills were identified with reference to the criteria shown in Table 2.2.

Basal till contains material which has been transported and deposited subglacially. The deformed and undeformed varieties of lodgement till (Hart and Boulton 1991) are not distinguished in this study. Clasts subject to active subglacial transport tend to undergo abrasion and crushing, resulting in the rounding, faceting and striating of clasts, and the production of fine-grained material. Deposits are consolidated by the pressure of the overlying ice, which can lead to a fissile structure, and subglacial deformation may occur resulting in folds and thrusts (Boulton 1976).

Meltout till is formed by the accumulation of material released by the melting of debris rich ice. This material may be of subglacial or supraglacial origin. In valley glacier systems, with an abundant potential debris supply from surrounding mountain slopes, it is likely that much passively transported supraglacial material will be present. This category includes both the meltout till and the supraglacial morainic till described by Boulton (1976) and Boulton and Eyles (1979). Meltout deposits are not consolidated, and usually lack clear internal structures, although there may be lenses and layers of meltwater worked material, or slump features due to the meltout of buried ice. They are usually poor in clays, silts and fine sands, as these are washed out by meltwater activity.

Table 2.2 Characteristics of glacial tills.

TILLS	BASAL	MELTOUT	GLACIGENIC SEDIMENT GRAVITY FLOWS
EXTERNAL FORM			
Shape of feature	fluted sheets	sheets, mounds	wedges, often irregular, smooth surface, lobate if high water content, may fill depressions and channels
Position w.r.t. glacier	subglacial	ice marginal	ice marginal
Dimensions of unit	laterally consistent	cm-2 m	1-4m
Other	lowermost glacial sediment, may be intruded into bedrock joints		uppermost glacial sediment
CLASTS			
Size	variety	variety	wide spatial variations
Shape	blocky, bullet shaped	variety	depends on parent material
Roundness	sub-rounded	angular to sub rounded	depends on parent material
Lithology	local derivation, may change with depth	local and distant derivation	depends on parent material
Surface features	striations, faceted, distal truncation	may have striae	depends on parent material
Orientation	long axes parallel to local direction of ice movement	long axes parallel or transverse to local direction of ice movement	weak, increases with water content, a axis aligned parallel or transverse to local flow direction
Imbrication	dip up glacier, tapered ends point up glacier	low dips, clasts parallel to plane of bed	may dip up-flow, or parallel to plane of flow
Sorting	poor	poor	possibly
Other	clast clusters and crushed clasts		
INTERNAL STRUCTURES			
Structures - primary	massive, or shear banding	massive diamicton, may be sedimentary banding	often structureless or chaotic diamicton, units may be stacked - angle of deposition decreases with water content
- secondary	fissile deformation structures	lenses and clasts of different material, draping over clasts, soft sediment clasts	flow lineations around obstructions, lenses and layers of well sorted silts and clays, deformed soft sediment clasts, sand and silt stringers, load structures
Support	matrix	matrix	mostly matrix
Joints, folds, shears	folds with noses streaked out, thrusts, boudins, subhorizontal anastomising joints	none	overturned or slumped folds, fold noses preserved, sheared structures, high angle faults
Contacts	sharp planar lower boundary with abraded bedrock	often interbedded with meltwater deposits and grades into flowtill	load structures, laminated silts on top, may be interdigitated with lenses and beds of sediment
Matrix size	silty, lots of fines	poor in silts and clays	wide spatial variations
sorting	poor	poor	possible - gravel layer at base
grading	no	may be inverse	normal or inverse
lithology	local derivation	variety	varies
compaction	high- platy	low if supraglacial debris	normal-low (Boulton - may be high)
density	high	low	usually lower than other tills
Lateral variations	low	low	high
Vertical variations	low	variable - may be high	may be high
Other	clast pavements at base, clast clusters on surface		may fill depressions and channels, or have erosive channellised bases

Sources: Boulton 1976, Boulton and Eyles 1978, Dreimanis 1988, Lawson 1988.

The final category of glacigenic deposit is flow till. Glacigenic sediment gravity flows result from the resedimentation of previously deposited glacial material down slopes, frequently in a water saturated form. They are particularly common in ice marginal situations where there is much meltwater, and where there are steep slopes. The constituent materials depend on those of the parent till, and the internal structures reflect downslope movement and washing and sorting by meltwater and valley side stream action (Lawson 1988).

Fluvioglacial outwash spreads

Fluvioglacial outwash was identified on the basis of location, morphology and internal sediments, where visible. Outwash fans are located in a proglacial, valley floor position, and other fluvioglacial deposits such as kame terraces are formed ice marginally. Currently active fluvioglacial outwash fans have steep ice contact slopes, which may have small push ridges, and gentle distal slopes of 12.5m km^{-1} close to the ice front, to 2m km^{-1} distally (Embleton and King 1975, Jurgaitis and Juozapavicius, 1988). Subaerial outwash fans may often have surface features such as palaeo-channels and kettleholes, which are depressions marking the location of buried ice that has subsequently melted out. Submarine outwash fans may be thicker (Boulton 1986), have a steeper distal gradient than sub-aerial fans, a deltaic structure at their proximal ends, and are less likely to have dead ice features on their surfaces. Outwash fans also have characteristic sedimentary structures. Close to the ice front of subaerial fans, there may be only weak sorting and stratification, and sediments may show internal deformations due to the meltout of buried ice. The clast size can be large, as meltwater streams issuing from the ice front can be flowing at high discharges, and incipient clast rounding may be evident. Coarse material may be found many kilometres from the ice front in former channels, whereas inter-channel areas contain fine sands. Distally, the materials become more sorted and stratified, cross-bedding structures become more common, clasts may become rounded, and the average grain size decreases. Often both small and large scale cross-bedding structures are found reflecting frequent channel migration over the outwash fan surface. In the proximal parts of the outwash a high, sometimes cyclic vertical variability in average particle grain size is common, reflecting the large and rapidly varying discharges over time associated with diurnal and seasonal changes in meltwater volumes. A fining upwards pattern may be detected due to retreat of the ice front. Sediment macrofabrics show clasts with the a-axis aligned parallel to the direction of transportation, and low dip New Para angles in an up-current direction (Jurgaitis and Juozapavicius 1988). Subaqueous outwash fans have large inclined beds, with the coarsest material concentrated close to the former ice front (Boulton 1986). An additional factor can be used to help identify outwash deposits dating from the LLS, where these are close to the sea. As sea-levels during the LLS were higher in the study area than they are at present (Section 1.4), the outwash was graded to this higher sea level, and has often been dissected by Holocene fluvial activity, leaving river terraces.

Striations and friction cracks

Striations are linear scratches on bedrock surfaces which are made by the action of subglacial clasts moving downglacier. Friction cracks can be of a variety of forms, but are usually crescent shaped fractures, with the horns pointing down-glacier. They are also formed by the movement of subglacial clasts. Crescentic gouges, by contrast, have horns that point up-glacier. There have been some recorded instances where the horns of crescentic gouges have pointed down-glacier. This is considered unlikely here, however, as Thorp (1981), in a study of hundreds of these gouges in an area immediately east of Loch Linnhe, found only six instances of reversed crescentic gouges, and these were all found in association with normally aligned examples. Both striations and crescentic gouges are commonly only <5mm deep at the most, and so their preservation is dependent on there having been less bedrock surface weathering than this since the end of the LLS. Previous work (Reed 1988) has suggested that glacial markings are poorly preserved on schists and gneisses, the most common bedrock types in the field area.

2.3 Results

A summary map of all the geomorphological evidence found is shown in Fig 2.1 (inside rear cover); more detailed maps of important areas appear later in the chapter.

2.3.1 Glacial drift

For the purposes of geomorphological mapping, glacial drift was taken to refer to deposits that are most probably glacigenic in origin, and occur as sheets mantling the base and slopes of valleys, or small sloping terraces on the valley sides. The thickness and extent of glacial drift deposits can be judged on the basis of stream dissection, banks and sections beside roads and tracks, and by the absence of bedrock exposures. Drift thicknesses observed in this manner were recorded in the field. The genesis of these deposits was assessed by examining any exposed sections and comparing the sediments present with the characteristics shown in Table 2.2. Most of these sections are found on the lower valley slopes alongside forestry and estate tracks, with some additionally in small quarries and tips, and in stream bank sections. In addition, in order to examine the regional distribution of drift, an east-west transect across the whole former LLS ice cap was selected, and all visible information as to the thickness of glacigenic deposits along it recorded.

Fig 2.1 shows that drift cover is most extensive in the northeast of the study area around Loch Eil and it's tributary glens, and cover decreases both to the south, and especially to the west. In particular, there is very little drift on the slopes around the western sea lochs, especially Loch Moidart and Loch Ailort, and the seawards sections of the glens feeding into them. There is also very little drift in the whole of the trough running west from the head of Loch Shiel through Loch Eilt to Loch Ailort, and in the upper Shiel trough. In individual valleys, drift deposits are concentrated in the mid and upper sections of the glens, and the basal parts of the glens, with deposits often commencing below the steeper glen sides above them, for example in Coire an Lubhair, Strontian Glen and Glen Tarbert. Till cover, often thick, is also common in hollows or basins on the sides of glens, immediately down-valley of rock bars and spurs, in side valleys which trend at a high angle to the main troughs, and in the lee of some cols.

There are three main types of sediments in glacial drift, illustrated in Fig 2.2, and they are often found in close association with one another.
An example of the first type of sediment is shown in Fig 2.2a from Glen Tarbert (NM 876 601). This section shows a loose, uncompacted diamict with a sandy and gravely matrix, angular unsorted clasts (Fig 2.3), and no large scale internal structures. In other sections small lenses and layers of partly sorted and stratified sands may be contained within this type of sediment, and these may have distorted bedding structures. Clasts in this type of sediment vary between platy and blocky in shape, and the upper three graphs of Fig 2.3 show that the clasts are mostly very angular to subangular. The first graph in Fig 2.4 illustrates the predominance of course grain size fractions in the matrix. This type of sediment is widespread. In two cases, north of lower Loch Shiel (NM 701 704) and Glen Doilet (NM 823 677), these till deposits have pods of soft sediment contained within them. This soft sediment is commonly deformed, for example into stringers, and is composed of a finer grained more consolidated diamicton.

Fig 2.2 (overleaf) Plates of sections showing different types of till.
a. Type 1 - meltout till, b. Type two - flow till, c. Type three - basal till.
See text for explanation.

A

B

C

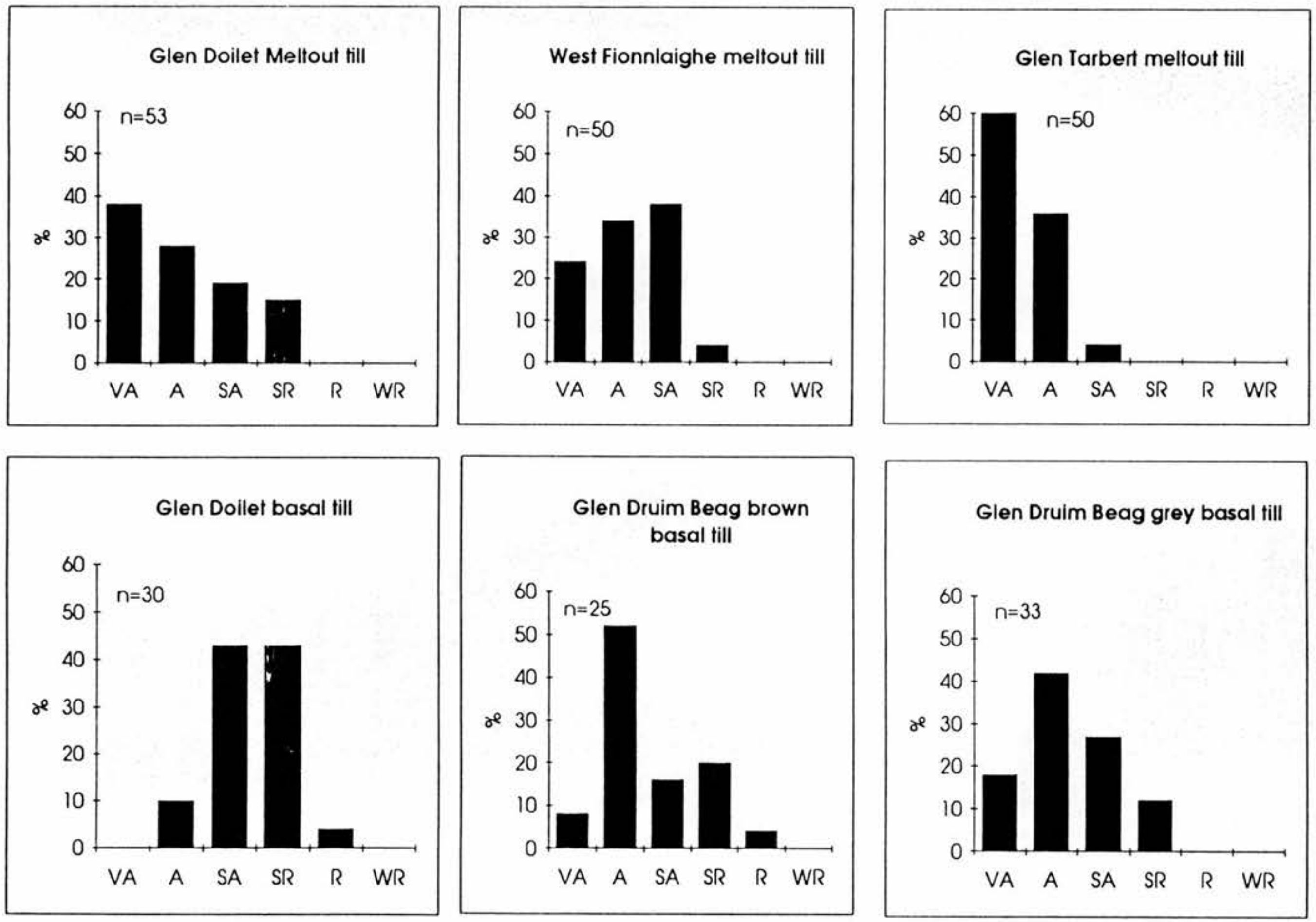
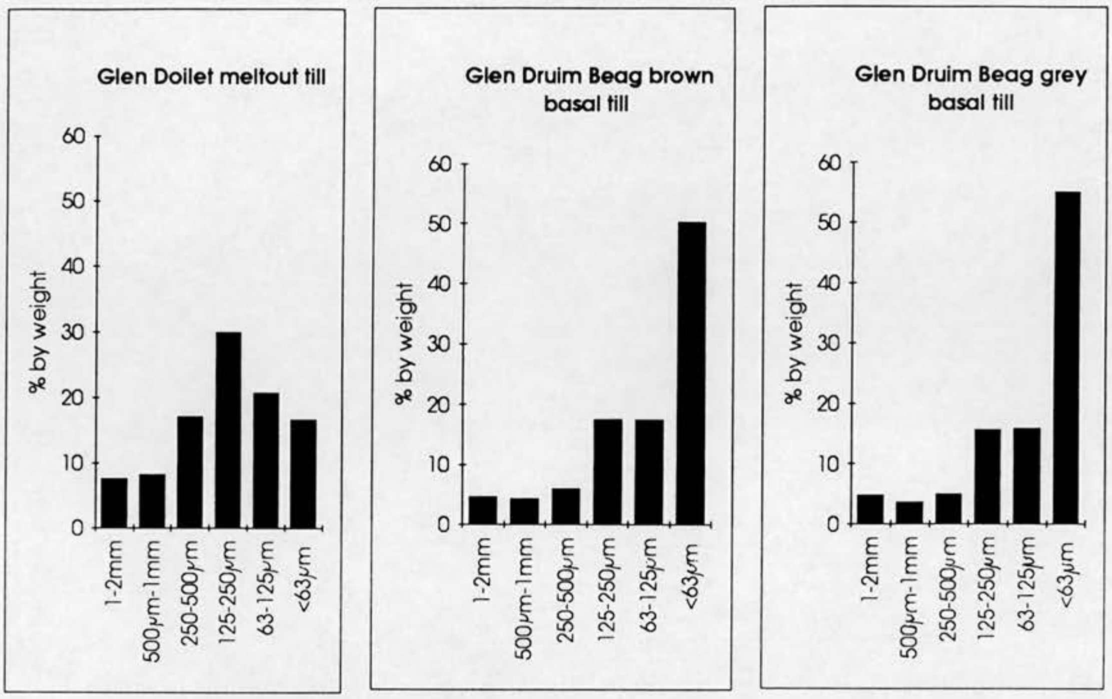


Fig 2.3 Angularity of clast (>8mm) samples from six tills



N.B. <63µm includes loss on wet sieving

Fig 2.4 Grain size distribution of matrix (<2mm) of three till samples

The second type of sediment is also widespread, and similar to the first, but with some degree of sorting and weak stratification. The beds may dip downslope, down valley, or a combination of the two. Clasts are similarly oriented, and they may include some which are subrounded. There may be some small drape structures of fines over larger clasts. An example of such sediments is shown in Fig 2.2b from Cona Glen (NM 983 716). This section shows weak downslope bedding with some clast concentrations at the base of units. Types one and two are commonly found in close association with one another, and separate units are not easily distinguished.

The third type of sediment is a consolidated diamict and an example in Glen Druim Beag (NN 001 803) is shown in Fig 2.2c. This section shows an upper brown till and a lower grey till. Both have clasts set in a consolidated matrix of silts and clays. The clasts in this type of sediment are commonly blocky and sometimes bullet shaped with striations, and Fig 2.3 shows that these clasts are consistently less angular than those from tills of type 1. These clasts are set in a consolidated, platy matrix of silts, clays and sands which may be faulted and jointed. This matrix contains a higher percentage of fines than the matrix of type 1 sediment (Fig 2.4). Type 3 sediment was mostly observed in small, partly degraded sections, but better exposures are found in Glen Doilet and north of Loch Eil in Glen Druim Beag (Fig 2.2c). Where visible, this sediment is often underlain by bedrock, and often overlain by types 1 and 2 deposits, for example in sections in Glens Scaddle, Fionnlaighe and Suileag. In most cases this type of deposit is brown in colour, and of very limited thickness, often just infilling depressions in the underlying bedrock surface. In both Glens Doilet and Druim Beag, however, where larger thicknesses are present, two diamictons of different appearance can be distinguished. The upper part is brown and the lower part grey (Fig 2.2c). In Glen Druim Beag the contact between them is sharp, whereas in Glen Doilet it is less clear, but in both cases pods of the lower till are contained within the upper. In Glen Druim Beag the concentration of clasts is much higher in the lower diamicton, and the average clast size is also much smaller than in the upper one.

The first type of sediment is interpreted as a meltout till containing debris of a mainly supraglacial origin (c.f. Table 2.2). This is characterised by angular, platy shaped clasts, and a sandy, massive, uncompacted matrix. The two samples from meltout tills both contained a high proportion of very angular and angular clasts, similar to those described from supraglacial debris by Benn and Ballantyne (1993). The sample from Glen Tarbert is in aggregate more angular that from Glen Fionnlaighe; this may be explained by the fact that the quartzite bedrock in Glen Tarbert is resistant to edge-rounding. It is also possible that the meltout till in Glen Fionnlaighe may contain subglacially or fluvioglacially transported material in addition to supraglacial debris (cf Dreimanis 1988). The lenses and layers of fines may represent sorting by the action of meltwater. Distortion of these units is due to disruption by the melting out of buried ice. The soft sediment rafts found in these deposits are interpreted as resulting from the incorporation of subglacial sediment rafts into an englacial position.

The second type of sediment is interpreted as a deposit of a mostly supraglacial meltout origin, which has been subsequently reworked by a combination of meltwater, valley side streams, and sediment gravity flows. It is similar in composition to meltout till, but shows some degree of downslope or downvalley internal stratification.

The third deposit is interpreted as a basal till containing material of subglacial origin. This is typically a compacted diamicton with a matrix of silt and fine sand, and sub-angular to sub-rounded, blocky clasts. Fig 2.4 shows that the basal till samples contain clasts that are less angular than the meltout tills, although that from Glen Druim Beag is more angular than is typical (Benn and Ballantyne 1993, 1994, Dowdeswell 1986). This may reflect geological characteristics or limited amounts of active subglacial transport. The apparent presence of two basal till units in Glen Doilet and Glen Druim Beag may reflect deposition during two different glacial episodes, a weathering difference, a local readvance, the difference between deformed and undeformed lodgement till or a change in subglacial conditions during the same glacial event.

In a few locations drift deposits have been found with different internal sediments. One of these is on the east flank of mid Gleann Dubh Lighe. Here, numerous trackside sections show up to 1m of sands with contorted bedding, overlain by a diamicton with weak downslope internal stratifications which is interpreted as a flow till deposit (NM945 818). These contortions are interpreted as representing dewatering structures in the bedded sands which were deposited into a body of standing water. Local and short-lived lakes are common around glacier margins.

The location of estate tracks mostly along the lower slopes of glens, and sometimes up into corries, means that sedimentary information has been primarily from these locations, with a dearth of evidence about the sediments on the glen floors and from glens with no tracks. Meltout and flow tills are the most widely distributed. Most of the basal till is located in the north and east of the field area. However, this may reflect the concentration of forestry tracks in plantations on the relatively gentle slopes in these locations.

In some places there are particularly thick drift deposits of 4m or more (Fig 2.1). These include several parts of the glens around Loch Eil, Gleann Sron a'Chreagain, lower Glen Finnan, parts of Gleann Dubh Lighe, Glen Aladale, Glen Fionnlaighe, upper Glen Guesachan, the Allt a'Bhuiridh and Allt an Utha glens, and the tributary glens to the south of Glen Tarbert and Cona Glen. The distribution of thick drift is likely to be under-represented on the map because afforestation and the limited presence of streams and tracks hinders detection. There are five locations where sections are found in thick drift deposits of more than 4m. These are in Gleann Sron a'Chreagain, at Ardgour, in Glens Aladale and Dubh Lighe, and in lower Glen Finnan. In Gleann Sron a'Chreagain (NN 053 728) the top 1.5m of a 7m high section are exposed in a river side slope failure. The sediments are an unconsolidated, partly stratified diamict, with angular to sub-rounded clasts set in a sandy, gravely matrix. The layers in the stratified parts dip downslope at an angle of 20°. This deposit is interpreted as till of a largely supraglacial meltout origin, which has been partially reworked by proglacial meltwaters and sediment gravity flows. The section in Glen Aladale (NM 832 778, Fig 2.5A) is at the base of a 10-14m thick till sheet in a steep sided narrow valley. It shows a diamicton with a loose sandy matrix containing clasts with an a-axis of up to 1m which vary from very angular to subrounded. There are some clast concentrations at the base of the section, and a crude internal stratification is evident with the angle of the beds roughly parallel to the local ground surface. This diamicton is interpreted as a flow till resulting from resedimentation of unstable glacial deposits on the steep slopes in this glen.

Fig 2.5 (overleaf) Plates of sections in thick drift deposits.
A. Glen Aladale, B. Glen Dubh Lighe. See text for explanation.

A

B

In Gleann Dubh Lighe (NM 937 813, Fig 2.5B) a small quarry section exposes a series of weakly stratified units which dip in a down valley direction. They consist of layers of fines and diamictons with poorly sorted clasts set in an unconsolidated matrix of sands and gravels. The clast sizes fine distally (left of plate), and in some cases are imbricated in a manner indicating deposition by flow from 80^{o} , which is upvalley at this point. These deposits are interpreted as ice proximal sediments which have been reworked by proglacial meltwaters and sediment gravity flows, possibly as part of a proglacial debris fan or ramp. Similar deposits are shown in a small quarry in lower Glen Finnan (NM910 814), which exposes 4m of ice proximal, reworked sediments overlain in places by laminated fines. The internal bedding in this section dips very gently downvalley and it is suggested that this results from sorting by proglacial meltwaters. On the west edge of the Corran outwash fan, south of Ardgour House, are a series of valley side cones, one of which has been excavated as a gravel pit (NM 992 638). The sections here show unconsolidated sands and gravels with various degrees of sorting and stratification. Some layers are trough cross bedded, and there is some evidence of slumped and distorted beds. The beds dip downslope, and in a clast rich bed, the clasts show a weak fabric dipping to the SSW. These cones are interpreted as ice marginal features containing glacial debris which has been sorted and resedimented by debris flows and valley side streams, and subject to slumping as ice support was removed.
Thick drift deposits are thus found along the floors of several glens, and in localised patches at the base of slopes. Available sedimentary evidence suggests that these deposits are either the result of downslope resedimentation of glacigenic deposits debris flows and slope wash, or have been transported in a downvalley direction by proglacial meltwaters or debris flows where glacial debris may have accumulated as a fan or ramp.

In some places the drift cover has a clear surface morphology. This can be in the form of a bench or terrace on the glen side such as on the north side of Gleann an Lochain Duibh (northern tributary to Glen Scaddle), and the north side of Glen Tarbert (see Fig 2.9). There are a series of 13 terraces in Glen Fionnlaighe. Some of these deposits are up to 25m thick, with surfaces that dip either downslope or downvalley and are sometimes dissected by stream channels. Those that dip downslope are usually closest to the glen sides and are bounded by linear ridges at their downslope edges. There are no sedimentary sections in any of the terraces. These features are interpreted as remnants of kame terraces banked up against ice marginal moraine ridges. The terraces closest to the centre of the glen with surfaces dipping downvalley are likely to represent material reworked by proglacial or ice marginal meltwater streams, or Holocene fluvial activity.
In other instances there is a steep rectilinear slope of linear or cuspate plan form in drift deposits. These latter are interpreted as former ice contact slopes, and are common on the

slopes above eastern Loch Eil, and in Glen Scaddle. These ice contact slopes record former ice marginal positions where there may have been a stillstand or other conditions allowing the accumulation of debris against the ice margin. This may accumulate in ice marginal lakes or by deposition from ice marginal streams. A trackside section in an ice contact slope in Glen Scaddle (NM 969 681) shows deformed beds of sands and gravels with some clusters of clasts, overlain by a massive diamicton with faint downslope internal stratification. The sands are interpreted as lake sediments which have been distorted by the meltout of supporting ice, and the clast concentrations may represent iceberg rafted debris. The overlying diamicton may be a flow till.
Much drift cover has been dissected by fluvial activity to create a hummocky surface. In some instances streams now occupy the channels formed and dissection is continuing at present. In other instances there are numerous channels which contain no contemporary water courses and have no obvious contemporary water sources. These are likely to have been cut by proglacial or ice marginal meltwater streams during deglaciation. These channels often run alongside ice contact slopes providing additional evidence for former ice marginal positions.

The glacigenic sediment thickness transect across Lochaber shows that the distribution of glacial sediments is uneven (Fig 2.6). Between Glenfinnan and the west coast there is little glacial sediment. From Glenfinnan to Corribeg on Loch Eil there are local patches of glacial sediments, and east of Corribeg as far as the Eastern LLS limits cover is almost continuous and usually at least 2m thick. Deposits are concentrated where there are outwash fans and moraines, particularly at the eastern LLS limit. East of this limit there is little glacial sediment around Loch Laggan, except where side glens join the trough, but east of the loch sediment cover is again fairly continuous with concentrations at ice marginal outwash terraces. There is thus a marked west - east contrast in the distribution of glacial sediments.

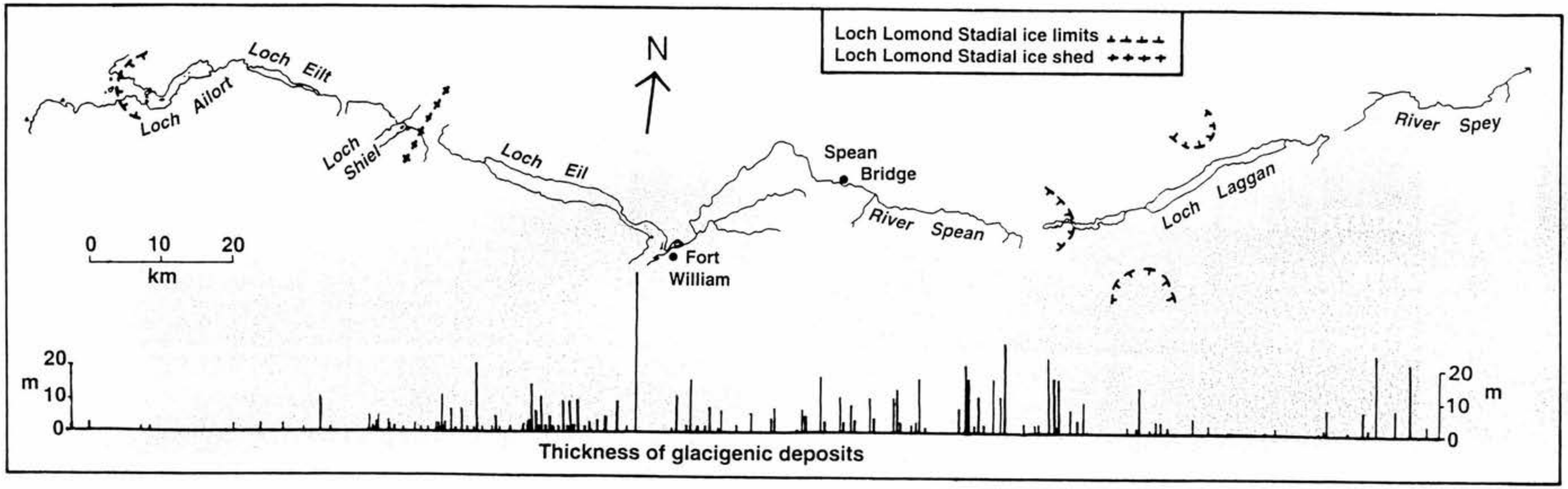


Fig 2.6 Glacigenic sediment thickness transect across Lochaber

2.3.2 Moraines

There are very few large moraines in Western Lochaber, but many glens contain hummocky moraine (Fig 2.1). These are usually small features 1-2.5 m high, although they may be up to ~10m high (Fig 2.7, highland cow on moraine summit indicates scale) and their surfaces are commonly scattered with boulders.

Fig 2.7 Hummocky moraines in Glen Tarbert

There are clear, isolated, small moraine ridges in many glens. Examples include those at Strontian village, in Glens Fionnlaighe, Finnan, Scaddle, Resipol, mid Glen Moidart, on the rock ridge west of Acharacle and at the mouth of the Coire na Cnamha, west of An Stac. Several of these morainic ridges are associated with bedrock ridges and bars. Others are at the mouths of side glens that join main loch troughs, for example a linear mound at the mouth of Glen Gour. This is composed at least superficially of gravels (R. McClean, proprietor, pers. comm.) and may be a moraine or ice-contact delta.

West of the mouth of Loch Shiel at Acharacle there is a low linear ridge 4.5m high and 300m long surrounded by up to 3m of peat. This is interpreted as a terminal moraine of the Shiel glacier (see Fig 2.15). In Resipol Glen a chain of small bouldery morainic ridges and fragments lie partly superimposed on a bedrock bar. These fragments enclose other small ridges and a drift limit (Fig 2.1) and are interpreted as the terminal deposits of a small glacier in this glen.

In several glens parallel moraine ridges and fragments are oriented transverse or obliquely to the strike of the glens, for example in glens Feith 'n Amean, Suileag, Cona, Strontian, upper Moidart, west Tarbert and Sron a' Chreagain. The best series of aligned moraine ridges are in western Glen Tarbert (Fig 2.8). Parallel alignments are developed over a wide area on the

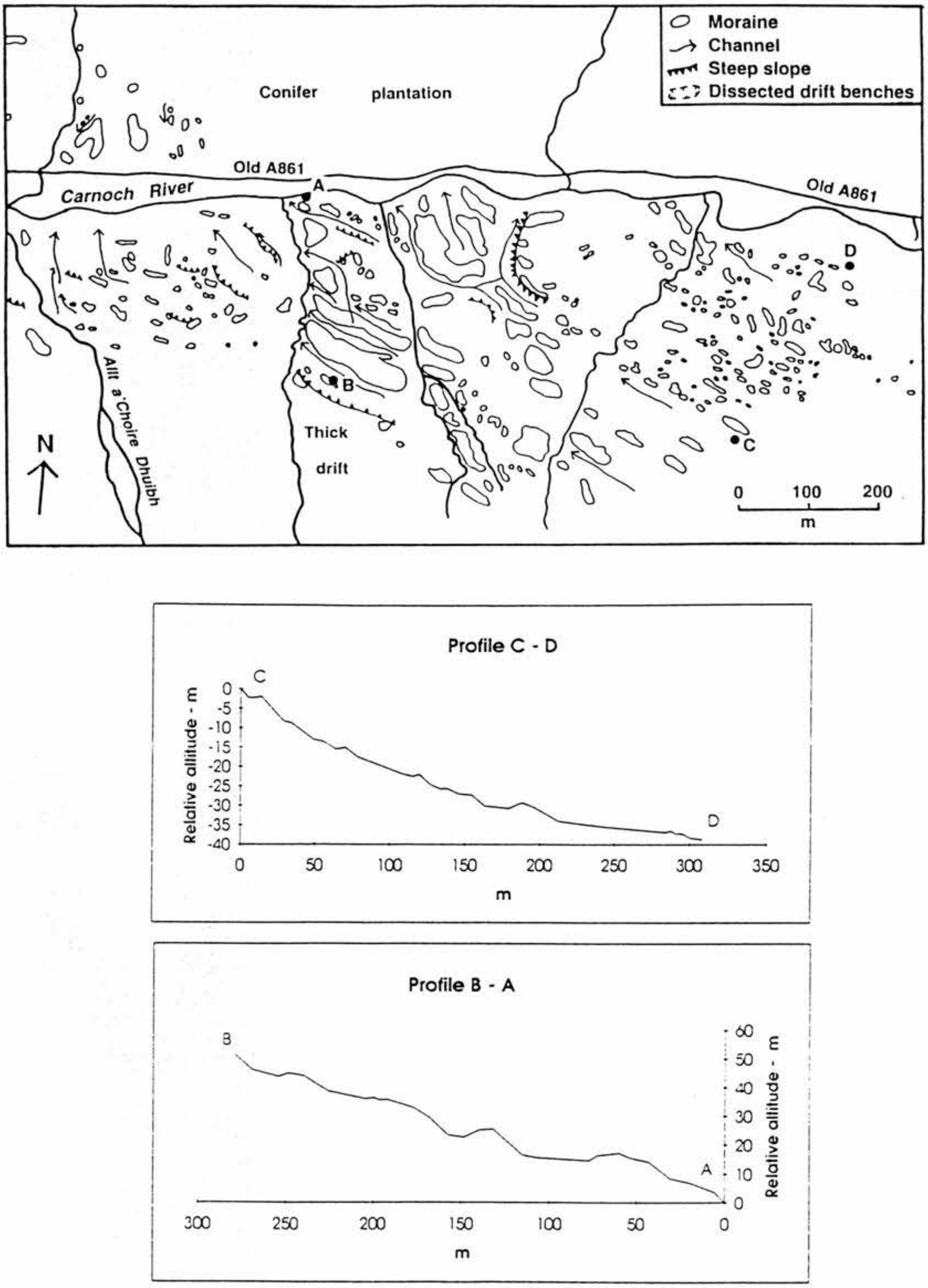


Fig 2.8 Aligned moraines in west Glen Tarbert

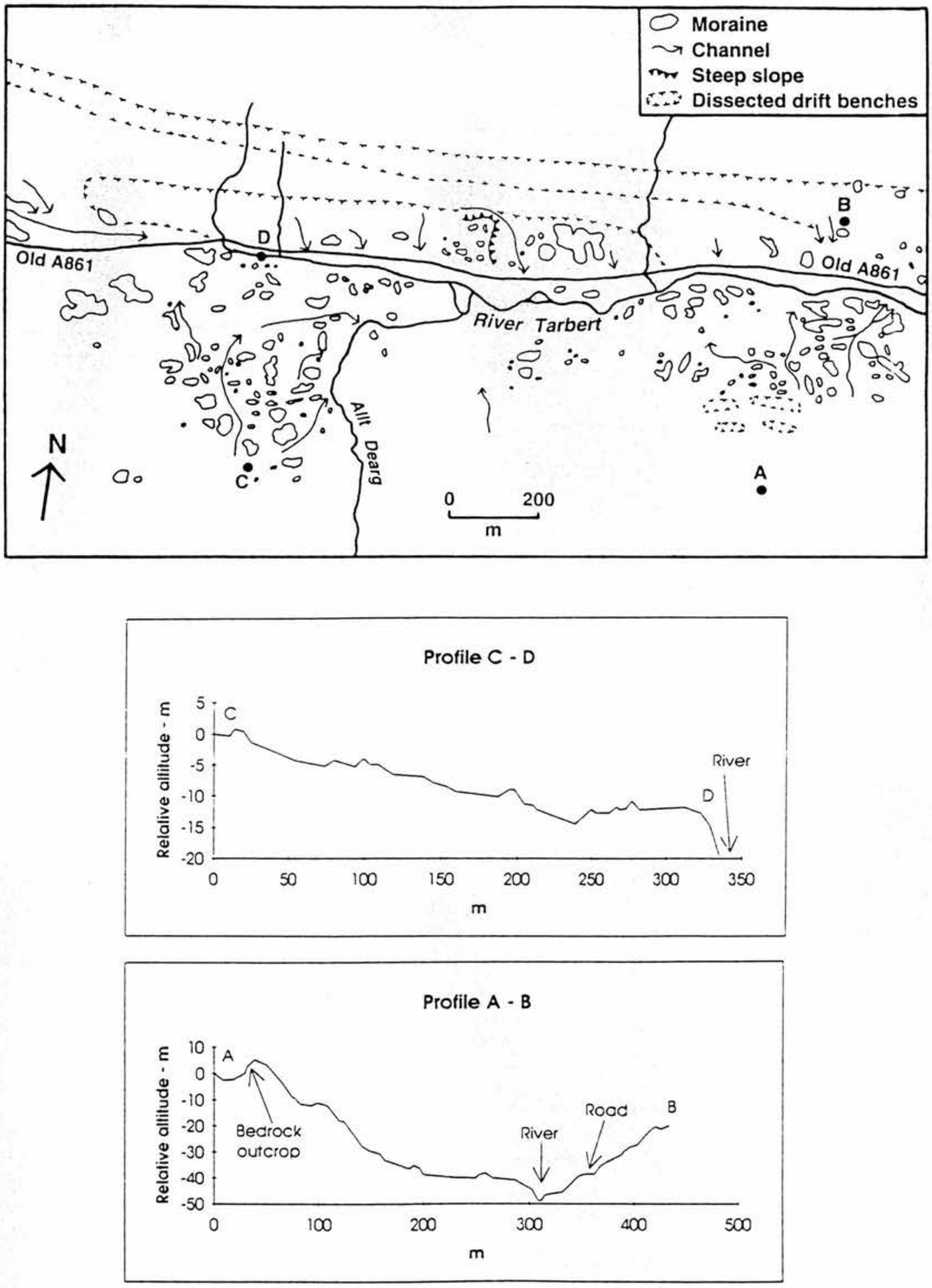


Fig 2.9 Chaotic moraines in east Glen Tarbert

south shores of Loch Eil on the high slopes above Blaich. Here, approximately 30 low-angled lineations can be detected at low light.
In some places there are isolated non-linear moraines, for example east of Loch Eilt, or areas of chaotically arranged moraines such as in Cona Glen, east Glen Tarbert, the head of Glen Scaddle, upper Glen Aladale, and mid Glen Suileag. Fig 2.9 shows a detailed map of a typical area of chaotically arranged moraine in Glen Tarbert.

There are just three good sedimentary sections in hummocky moraines in Western Lochaber, two of which are in linear moraine ridges. Each of these sections shows a diamicton of sub-angular and angular clasts set in an uncompacted matrix of sands and gravels, as shown in Fig 2.10 taken in Glen Fionnlaighe (NM 977 831). In places there are signs of weak downslope stratification (e.g. left side of Fig 2.10), and layers and lenses of fines (e.g. above and right of notebook in Fig 2.10). These characteristics suggest that the moraines are composed of supraglacial meltout material, which has been partly reworked by debris flows or the action of meltwater. None of these three sections show evidence of glacitectonic structures, or of subglacial till. All the trackside sections examined which cut through more ill-defined valley side mounds also show supraglacial meltout tills.

Fig 2.10 Section in hummocky moraine in Glen Fionnlaighe.

In two locations, Coire a'Bhuiridh and the northern tributary to Strontian Glen, there are low parallel aligned ridges. In the latter location these are oriented diagonal to the strike of the glen; in the former they are oriented parallel to the strike of the glen (Fig 2.11). In Coire a'Bhuiridh there are at least 100 longitudinal ridges, and these range from 0.5-12m high, 1.5-30m wide, and 10-500m long. The flute with the largest dimensions is on the right of the Plate in Fig 2.11(rucksack adjacent to boulder on extreme right of crest of this flute indicates scale). The air photograph in Fig 2.11 shows that the flutes are arranged in the corrie so that the ridge lines converge slightly from the back of the corrie to the mouth, where they are more parallel. Small stream bank sections show that the surficial sediments are a loose sandy diamicton containing angular clasts. This is interpreted as material of supraglacial origin, but is not necessarily typical of the sediments at depth within the moraines. At the back of the corrie the spatial distribution of boulders is such that they are concentrated on top on the ridges. One ridge appears to consist almost entirely of angular boulders, with an a-axis of up to 5m (left flute in Plate in Fig 2.11). Two lines of boulders continue from the back of this ridge to a crag on the back wall of the corrie.

The arrangement of these longitudinal ridges suggests that they are fluted moraines. Fluted moraines, however, are universally recognised as subglacial landforms, although they may have a mantle of supraglacial material deposited during deglaciation (e.g. Boulton 1976, Benn 1995, Bennett 1995). A model of subglacial formation may account for most of the flutes. However, it seems unlikely that a ridge composed of openwork angular boulders which is connected to a crag by a boulder line could have been formed in a subglacial position. It is likely that the bouldery flutes at the back of the corrie are each composed of supraglacial material which fell onto the former glacier from constant source cliffs and record former ice flow lines in a manner analogous to subglacially formed moraines. Interestingly, both these locations are situations where the LLS glacier would have been partly advancing upslope, which is where fluted moraines on Skye are located (Benn 1992).

Fig 2.11 (overleaf) Fluted moraines in Coire a'Bhuiridh.

A. Air photograph - South at top of plate.

B. Plate of flutes at back of corrie. Location shown on air photograph.


X

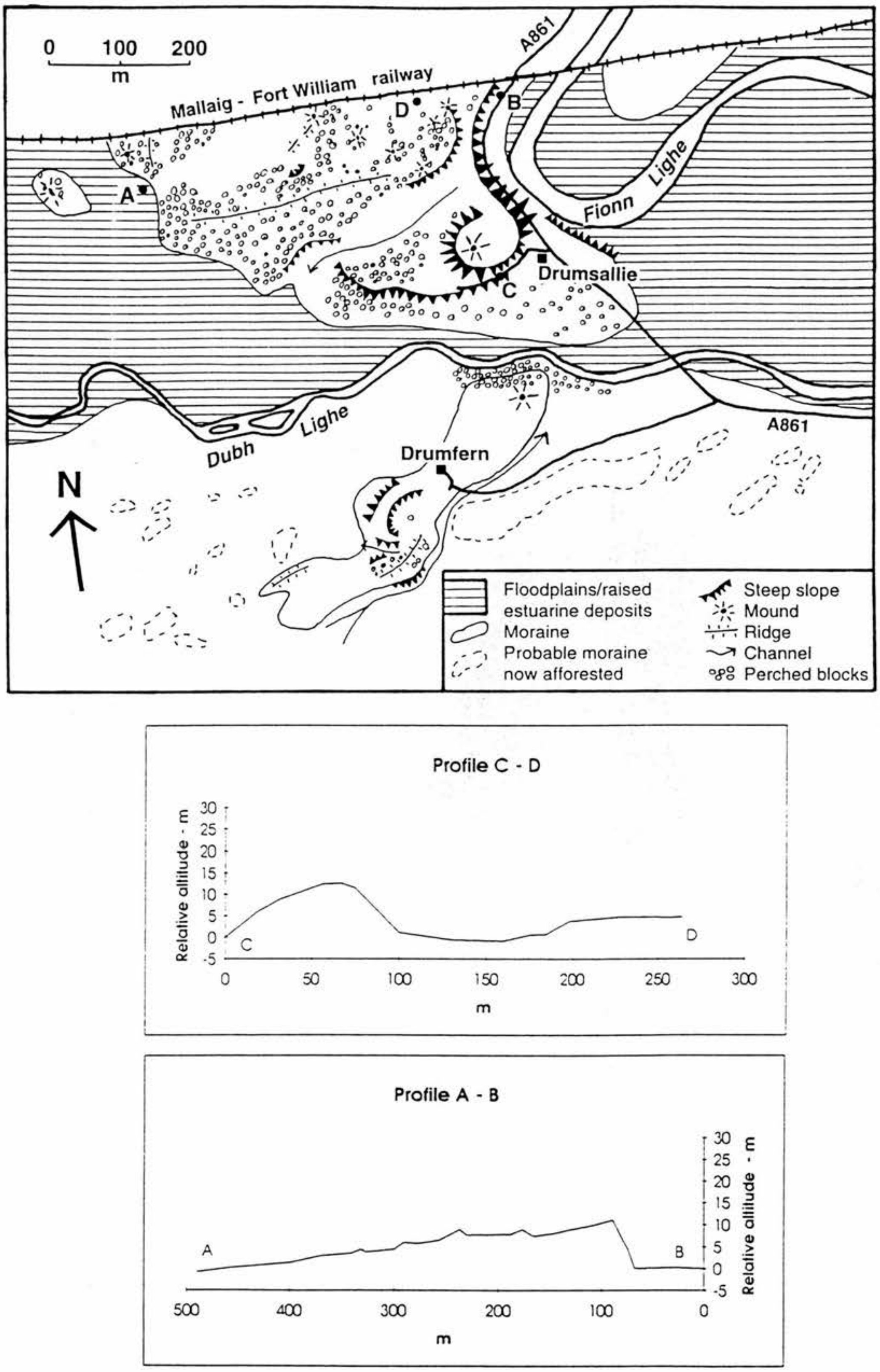


Fig 2.12 Loch Eil moraine

The largest moraine in the study area is at the head of Loch Eil. This consists of two parts, one to the north and one to the south of the river (Fig 2.12). The northern part is not a conventional ridge, but it can be clearly distinguished from the surrounding flat alluvial plains and raised estuarine deposits. The surface has both small and large scale mounds, and is dissected by channels. The surfaces of both are littered with boulders, in marked contrast to the flat plains around on which there are none. A section in the northern moraine shows an uncompacted diamicton with high lateral and vertical variability in internal structures. It is mostly composed of stratified sands and gravels with some distorted bedding, but close to these are massive diamictons, soft sediment rafts, and clast concentrations. The clasts vary from being sub-rounded to sub-angular; many are of a blocky shape, and some are faceted. The orientation and dip of clasts in one bed indicate deposition by flow from the north east. These deposits are interpreted as consisting of debris which has been transported supraglacially and, to a lesser extent, subglacially. They were deposited in an ice proximal location, some into a body of standing water, with the distorted laminations being due to the meltout of buried ice, and the clast concentrations resulting from debris flows and/or iceberg-rafted debris dumps.

Most of the moraines are in the upper parts of glens. The parallel, linear arrangement of some moraine ridges and fragments, and their asymmetric cross-profiles suggest that these are former ice marginal moraines recording the pattern of active retreat of the LLS glaciers. In two glens fluted moraines clearly show the final direction of ice movement, which, in one instance, was oblique to local topographic trends. Other hummocks are arranged chaotically. There are no large terminal moraines relating to the ice maximum in this area, unlike those found at the eastern margins of the Loch Lomond Stadial Scottish ice cap. There are, however, three small terminal moraines, all deposited at least partly on bedrock ridges. One marks the maximum of the Shiel glacier. Why?

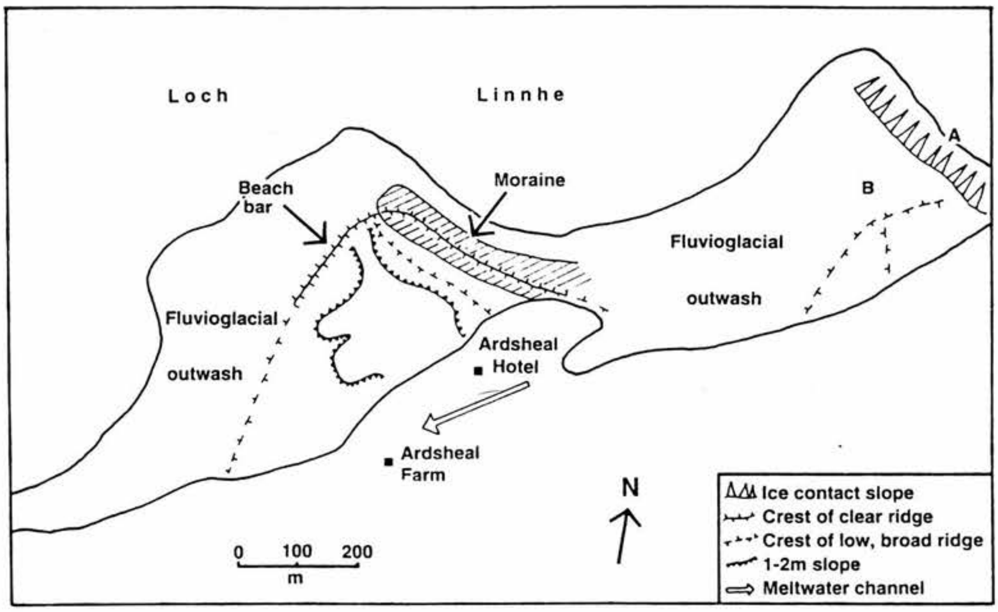
2.3.3 Fluvioglacial deposits

The distribution of fluvioglacial outwash in and around the study area is shown in Fig 2.1 and the characteristics of the outwash spreads are summarised in Table 2.3. There are large fans at Kentra Moss, Claish Moss, Mointeach Mhor and Corran. There are also significant spreads at Annat, Kentallen and Ballachulish. Fluvioglacial outwash fans are found only in close association with large tidewater or freshwater lochs. They are found at the mouths and heads of Lochs, at narrowings and shallowings, and where large side valleys join the lochs.

Table 2.3 Outwash fans

Location	Area (km^2)	Max height (m)	Gradient m/km	Surface Features	Observed depth of deposits (m)	Sections
Monaidh Mhor	6.25	19.39	3.35	ice contact slope, flat	>9.39	sands and gravels
Roshven	0.19	11.19	6.13	ice contact slope	/	none
Kentra Moss	6.19	16.78	5.98	kettlehole, eskers, ice contact slope	>3	cross bedded sands and gravels
Claish Moss	10.19	20.90	2.45	kettleholes	>6.23	sands and gravels
Aladale	0.19	>20.0	/	ice contact slope, hummocky surface	>12.64m	crude horizontal stratification, large clasts in clusters, sandy gravel matrix, clast support
Glen Laudale	0.64	17.89	16.4	flat	/	none
Strontian	0.66	>14.32	1.5	palaeochannels and moraines	>4	none
Inversanda	0.53	13.99	7.16	flat, dissected	>2m	clay bed, sorted gravels and cobbles
Kentallen (4) 1-3	0.06 0.57	13.3	1^{o}	possible kettlehole and channels, ice contact slopes	/	not clear, gravels and cobbles
Corran	4.0	27.60	4.80	dissected, channels, kettleholes	>24	well sorted beds of sands → cobbles
Onich	0.5	14.64	/	slopes seawards, beach bar?	/	surficial raised beach sections
Ballachulish	0.81	14.64	8.92	channel or ketttlehole?, beach bar	/	poor
Caolasnacon (Loch Leven narrows)	0.19	21.28	17.69	dissected	>5m	well sorted beds of clays → cobbles
Inverscaddle	1.45	12.23	5.58	dissected	>5	well sorted beds of clays → cobbles
Annat	0.72	12.48	/	kettlehole	>45	beds of sands, gravels and cobbles overlying fine sands and silts

The outwash fans all have low surface gradients, and some have steep ice contact slopes (see Fig 2.13). Some of the larger spreads have surfaces features such as kettleholes and channels, which indicate sub-aerial deposition. One fan at Kentallen has a ridge along the proximal crest, which may be of morainic origin (Fig 2.13). Another fan, at the Roshven narrows on Loch Ailort, has been previously interpreted as a 'high' lateglacial raised beach (Dawson 1989, 1994a). As this fan has a clear proximal ice contact slope and gentle distal slope, a proglacial outwash origin is favoured here.



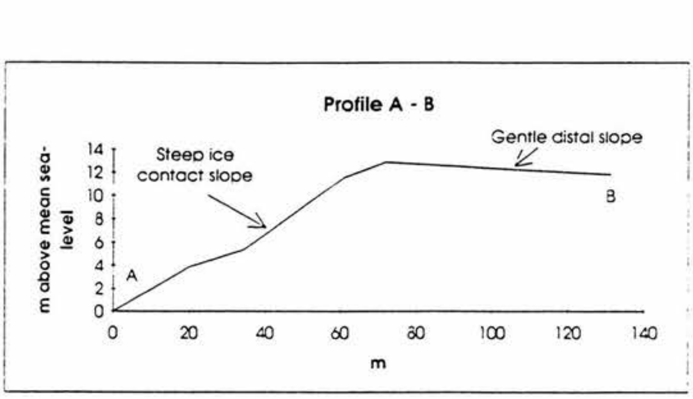


Fig 2.13 Geomorphology of Kentallen outwash fans 3 and 4

Sediments are exposed in several of the outwash fans. A small section close to the north end of the deposits at Corran (NN 013 633) shows horizontally bedded, weakly sorted subangular gravels and cobbles, and a section further west shows weakly stratified to massive sands, gravels and cobbles. Small sections near the coast at the centre of the Annat fan (NN 076 767) also show stratified sands and gravels. These are poorly sorted and mostly sub-angular. Bores sunk prior to the construction of the Annat pulp mill showed up to 12m of coarse sands and gravels, sometimes overlying fine sands and silts. Rockhead was not reached at 45m in the deepest bore. Sections in smaller spreads of outwash at Caolasnacon (NN 138 611), Aladale (NM 823 748) and Inverscaddle (NN 017 689, Fig 2.14) all show weakly stratified horizontally bedded sands and gravels, containing mostly subangular or angular clasts (see Fig 2.17) which are up to 1m in a axis. The sediments show marked vertical variations in clast and matrix grain size and degree of sorting, varying from small layers and lenses of silts and clays to clast supported cobbles (e.g. next to trowel in Fig 2.14). All these sediments are typical of ice-proximal fluvioglacial deposits. The large clast sizes and poor sorting suggest high energy deposition, and the marked vertical changes reflect the widely varying discharges associated with glacial meltwater flow.

Fig 2.14 Inverscaddle outwash section

The outwash at the mouth of Loch Shiel has a prominent suite of surface features (Fig 2.15). There is a clear ice contact slope of cuspate plan form, eskers, and a kettlehole (Peacock 1971). The surface of the Kentra outwash fan is at a maximum at the top of the clear ice contact slope, and grades away westwards for 3.5km both from this slope (Shennan et al. 1994b, Dawson 1994b), and from the efflux of a meltwater channel through the rock bar (Wain-Hobson 1981). This meltwater channel is close to the inferred terminal moraine

(Section 2.3.2) on this rock bar. The surface slope of the outwash suggests that the fan was developed when the ice terminus was at, or close to, the positions marked by the ice contact slope and the moraine.

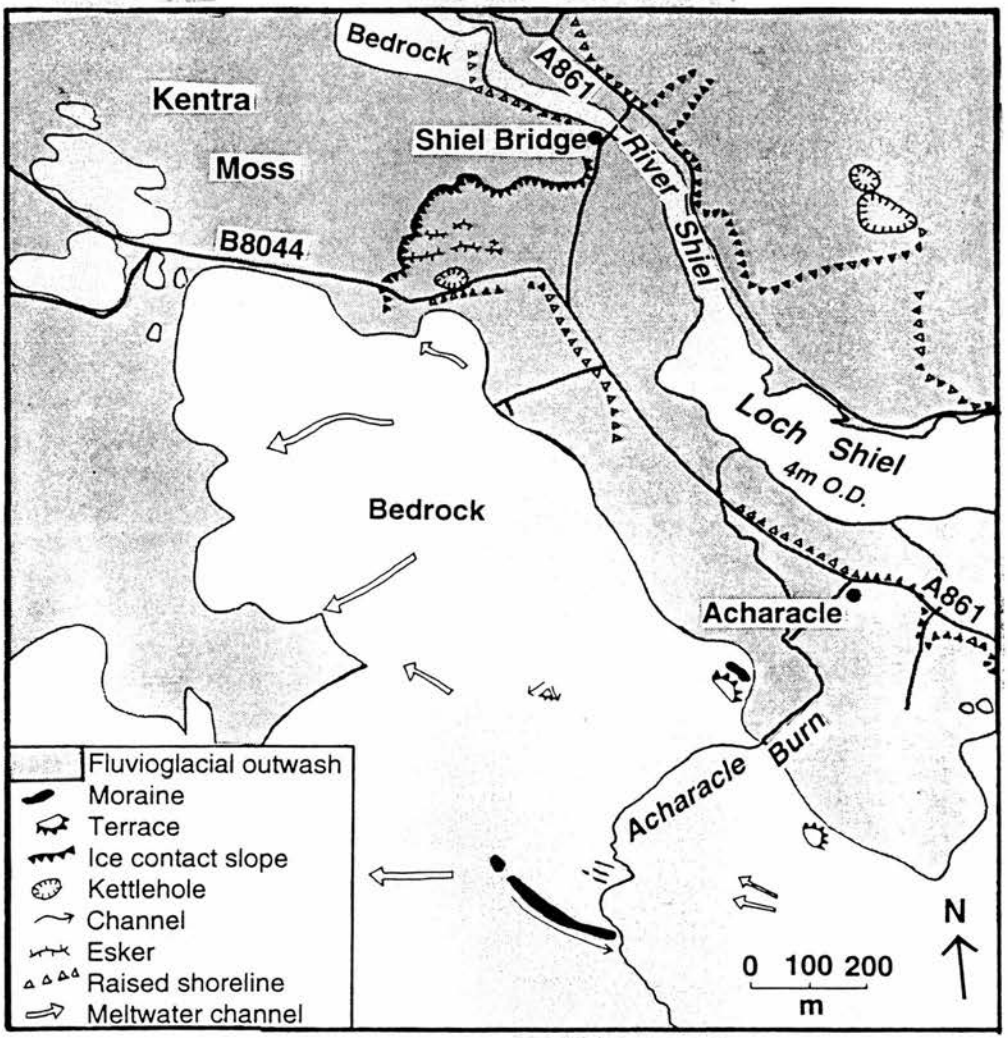
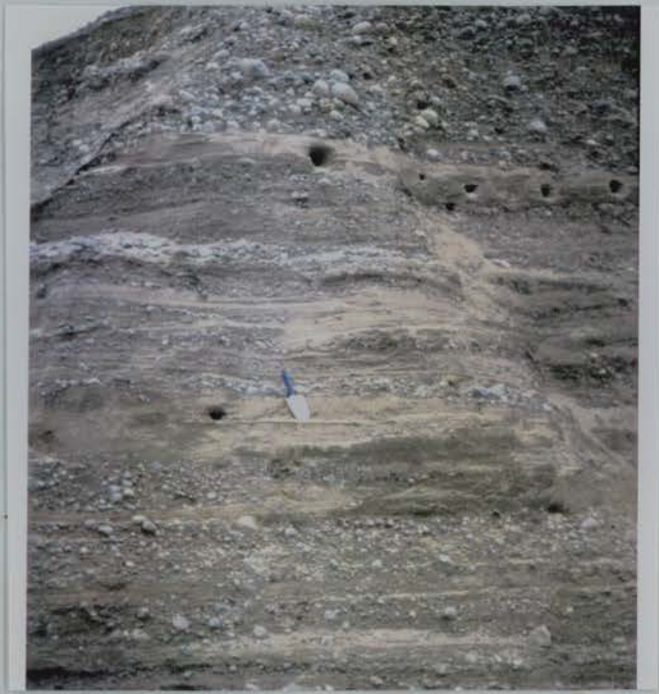


Fig 2.15 Glacial geomorphology of the Acharacle area

Fig 2.16 (overleaf) Plates of sections in Kentra outwash, showing transition from proximal to distal sedimentation. A. Shiel bridge B. Arivegaig quarry C. Arivegaig sream (knife indicates scale).

A



B

C

There are several sections in and around the Kentra outwash. At Shiel Bridge (NM 678 691) there is a large sand and gravel quarry exposing a 3-6m x 200m section around three sides of a square. These exposures show sub-horizontally and cross bedded sands and gravels containing mostly sub-rounded clasts of up to 15cm in diameter (Fig 2.16a). The sub-horizontal beds dip at an average of 1^{o} to the west. There is no sign of post- depositional disturbance. The clasts and matrix vary from being poorly to well sorted. The large clast sizes and incomplete nature of the sorting suggest that these deposits were laid down by proglacial meltwaters fairly close to an ice front. Three miles west of this section there is a small gravel pit at Arivegaig (NM 659 679). This exposes well sorted sands and gravels dipping at 9.5^{o} to the WSW (Fig 2.16b). The clasts are mostly subrounded and of up to 7cm in a axis. These overlie well-sorted laminated sands, the contact being at 6.64 m OD. The gravels are interpreted as distal fluvioglacial outwash deposits deposited in a channel bar in a meandering channel system on the outwash surface.
Still closer to the sea is a third section (NM 655 683). This shows well-sorted sets of cross-bedded sands and silts, with some channel features, and overlain by a few gravel beds (Fig 2.16c, knife in centre indicates scale). There are also isolated concentrations of larger angular clasts of a-axis up to 20cm, truncating the laminations in the sands below. In addition there are some deformed soft sediment rafts of a grey silty clay which are separated from the sands below by an erosional unconformity (centre of plate). These are all overlain by horizontally-bedded sands with some clasts. The sands show some ripple bedding. The sands are interpreted as ice distal deposits, probably lain in a subaerial or shallow tidal environment. The larger angular clast concentrations could be interpreted as dumps of iceberg rafted or fluvially transported debris, and the clays as rafts transported by sea ice (Dawson pers. comm.) or fluvially, originating from tills or glacimarine deposits upglacier. The sedimentary evidence available at different points on the Kentra outwash thus shows a westwards transition from ice proximal deposits around the clear ice contact slope to ice distal deposits, supporting the interpretation that the outwash was deposited while an ice margin lay at the ice contact slope at Acharacle.

Above Dalnabreck is a clear ice marginal fluvioglacial feature (NM 702 704, Fig 2.1). It is a large, almost flat-topped deposit at 80-90m OD, located at the south end of a meltwater channel which cuts across the ridge between the Shiel and Moidart troughs. It has been dissected to some extent on the west side by a small stream. There are a few small sections in this slope which show beds of well sorted, subrounded sands and gravels. These have been erosionally truncated in one place by a channel shaped fill of clast supported cobbles and gravels. Most of the laminations dip at approximately 23^{o} to 220^{o}. This feature is interpreted as a deltaic deposit formed in an ice marginal lake. Lithology counts suggest that the debris source was probably from the Shiel basin, rather than via the meltwater channel from the Moidart basin (see Section 2.3.5).

Fig 2.17 shows the clast roundness of samples of 50 clasts from 10 of the outwash fans in Western Lochaber. These measurements from outwash fans were made in order to try to reconstruct the transport histories of the clasts involved. In general, clasts which have undergone little active transport, such as frost shattered debris and supraglacial material, are platy shaped (low c:a axial ratios) and have a high proportion of very angular and angular clasts. Subglacially transported clasts, by contrast, tend to be predominantly blocky shaped (high c:a axial ratios), and subangular to subrounded (Benn and Ballantyne 1993, 1994). Aggregate clast shape and roundness indices of the same samples shown in Fig 2.17 are listed in Appendix 1. Clast shape appeared to be strongly influenced by clast geology, with unfoliated, isotropic rocks producing preferentially blocky clasts, whereas laminated, bedded or foliated rock types produced platy clasts. This is in accord with principles discussed by Boulton (1978), and clast shape is not analysed further.

Fig 2.17 shows that the outwash deposits in this area contain a wide variety of clast roundnesses; there are wide ranges within samples, and between fans, and smaller ranges within sites and within outwash fans. Those from the small outwash fans at Inverscaddle, Caolasnacon and Aladale all contain a high proportion of subangular clasts, with the sample at Inverscaddle containing mostly angular clasts. The two samples from each of the two fans at Annat and Corran show that the clasts in different parts of each fan are not of uniform characteristics. Most of the clasts in all of these samples, however, are subangular and subrounded. Comparison with the angularity of the meltout and basal till clast samples shown in Fig 2.3 suggests that these deposits may consist of reworked supraglacial and/or subglacial material which was subject to limited active transport in the fluvioglacial system (c.f. Tucker 1981, Boulton 1978). In all these cases the variety of clast angularities suggests that the clasts have undergone a range of transport histories.

The clast samples from the sites in the lower Shiel basin (the lower seven graphs in Fig 2.17) show aggregate clast characteristics more rounded than those from the other outwash deposits, although usually with large amounts of internal variability. The three samples taken from different points in the same quarry at Shiel Bridge are similar in that they all contain more subrounded clasts than any other category, but there are differences in the amounts of subangular and rounded clasts. Some of the samples in this trough contain clasts that are more rounded than samples from both basal tills (Fig 2.3) and rivers in Western Lochaber and are of similar roundness to beach samples (Aggregate clast shape and roundness indices from river beds and beaches of high and low fetches in Western Lochaber are listed in Appendix 1). This suggests that the clasts have undergone large amounts of active transport. This may have

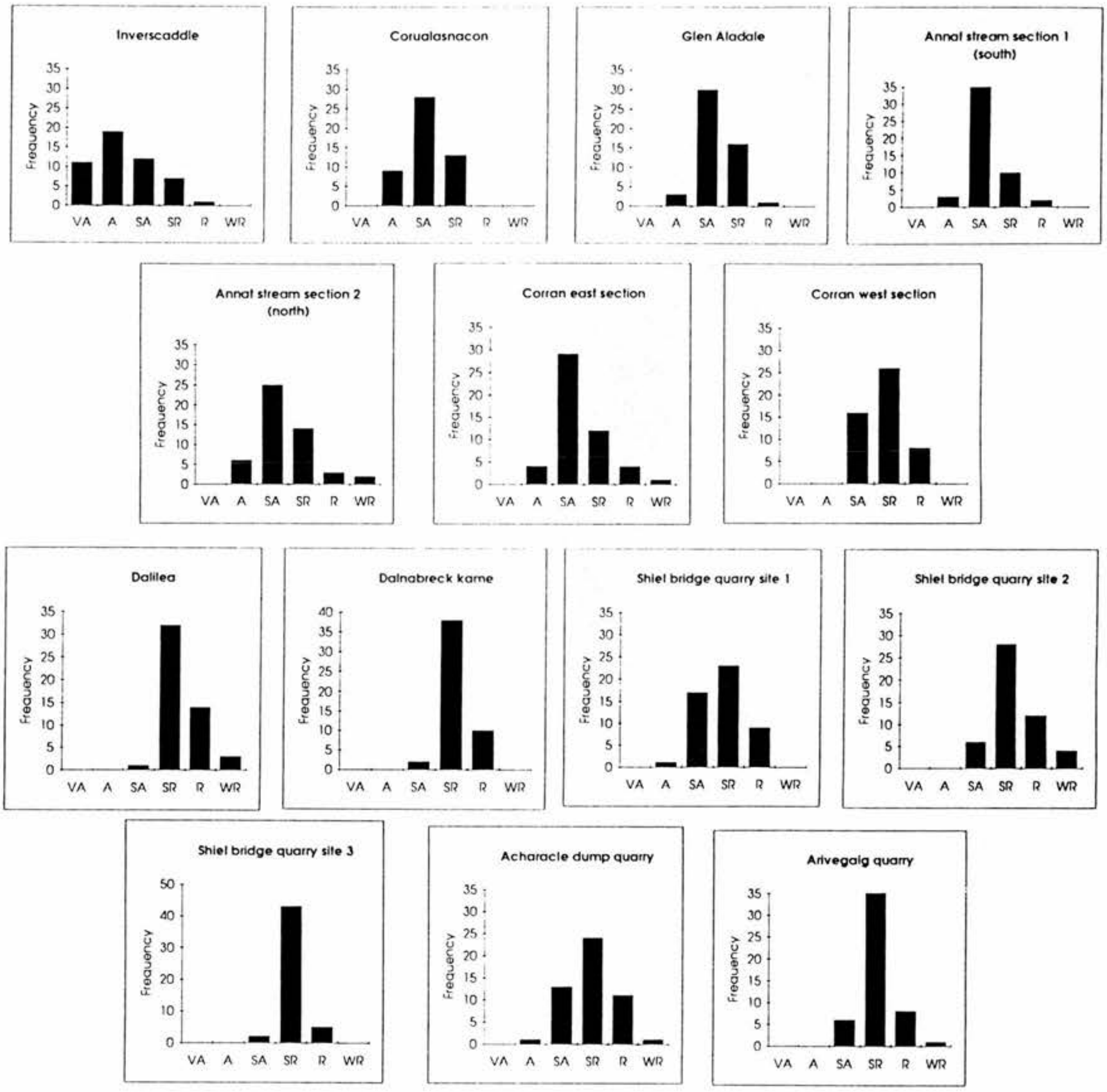


Fig 2.17 Angularity of clast samples from fluvioglacial outwash

occurred in LLS and/or previous subglacial or fluvioglacial transport systems, and/or during previous fluvial or beach activity. Further measurements of clast shape and roundness using samples of constant lithology would be useful for investigating these issues more fully.

Most of the fluvioglacial deposits in the area are proglacial outwash spreads located on the shores and at the mouths of the Lochs in the area. Some show steep ice proximal slopes which record the position of the ice front at the time of deposition, for example at Loch Shiel, Corran, Ballachulish, Annat and Roshven. These are all places where the deposits are in the middle of glacial troughs. Outwash spreads at the heads of the lochs or at the mouths of side valleys, do not usually have ice contact slopes, except in Glen Aladale. Many of the outwash fans at altitudes above ~12m show evidence of sub-aerial deposition. There are a few, mostly small, examples of ice marginal fluvioglacial features.

2.3.4 Concentrations of perched blocks

Most of the slopes in the study area are scattered with boulders, especially the surfaces of moraines, but there are some additional areas where particular concentrations of perched blocks were noted. The main concentration is north and east of the head of Loch Sunart where the boulders are concentrated on the northern valley side and the plateau above. They are most prevalent in a band through a zone of hummocks, many of which are bedrock features. These zones do not exactly coincide, however; the boulders are present in lower densities west and east of the hummocks, and the hummocks are present west of the boulder band. The blocks are large, averaging a-axes from 0.5-3m. Examination of 175 blocks on a transect from the plateau to the valley floor showed that 96% were of Strontian granite, and just 4% of Schist. The distribution of this boulder band closely follows the outer limit of the Strontian granite intrusion. It is provisionally concluded therefore, that Strontian granite is easily eroded into large blocks, and that the location of this boulder spread is geologically controlled. This is in contrast to Wain-Hobson's (1981) conclusion that the extent of the boulders marks the maximum extent of westwards flowing Loch Lomond Stadial ice. If so, it would not have been flowing over granite, but schists.

Boulders are also concentrated in upper Cona Glen, and at a number of cols. These include those north of Sgurr Dhomhnuill, East of Meall Mor, east of Meall a' Phubuill, and both north and east of Rosh-bhein.

2.3.5. Stone lithology counts

Most of Ardgour and Moidart is underlain by schists and gneisses, with various outcrops of igneous intrusions (Section 1.3, Fig 1.4). A few of these intrusions are readily identified in the field, and have a limited spatial extent, and are thus suitable for the investigation of ice transported erratic trains. Furthermore, there is a belt of granitic gneiss which is roughly followed by the present watershed in Ardgour (Fig 1.4). Examination of stone lithologies in deposits at the mouths of the glens running east and west from this watershed can give information as to whether the ice shed lay east or west or coincided with this belt of granitic gneiss. In interpreting information on glacial erratics it is important to recognise that there may be erratics transported from elsewhere during former glaciations, which have been incorporated into LLS deposits, and may thus not reflect the direction of transport by LLS ice. As previous icesheets are believed to have flowed in a westerly direction over the field area, with the probable exceptions of ice streams flowing down Lochs Linnhe and Shiel, potential erratic sources from the east are most likely to be involved.

A lithology count of a sample of 50 clasts was made in the field from near the mouth of each glen. These were from sections in moraine or drift where possible, or else from river and stream banks. Where the lithology could not be identified, a sample was bought back to Edinburgh for further identification. The results are listed in full in Appendix 2, and the implied ice flow directions are shown in Fig 2.18. This figure was constructed by drawing arrows down troughs to the sites where erratics occur from the closest point at which the lithology concerned outcrops at the surface (see Fig 1.4). The results shown in Fig 2.18 indicate that the Gabbro from eastern Ardgour was moved east, Strontian granite was carried west, and the Appinite outcropping between the Shiel and Moidart troughs was transported down both these valleys. Pegmatite appears to have been transported down the Shiel trough, down the Eil /Linnhe basins, and south eastwards across northern Ardgour. There are likely to be small local outcrops of this intrusion throughout much of the study area, however, so it is not possible to identify glacial transport directions from the dispersal of pegmatite erratics with certainty.

Granitic gneiss from the north of the area was carried both west to Loch Ailort, and East down Loch Eil. In northern Ardgour a little granitic gneiss was transported east, but this was not so in Glen Gour to the south. The high values in Glen Finnan and Glen Tarbert (see Appendix 2) are anomalous and not indicated on Fig 2.18. It proved difficult to distinguish granitic gneiss from psammitic gneisses in the field, and large outcrops of gneiss were noted in these two glens, which may have confused the picture. Micaceous semipelitic and pelitic schist from west of the Ardgour watershed was carried eastwards into Gleann an Lochain Duibh, but not into the other glens in Eastern Ardgour. Lithology counts from east of Loch Linnhe showed a consistently different suite of rock types to those found in Ardgour and Moidart (Appendix 2), suggesting that ice flowed into this trough from the mountains to the east as well as the west.

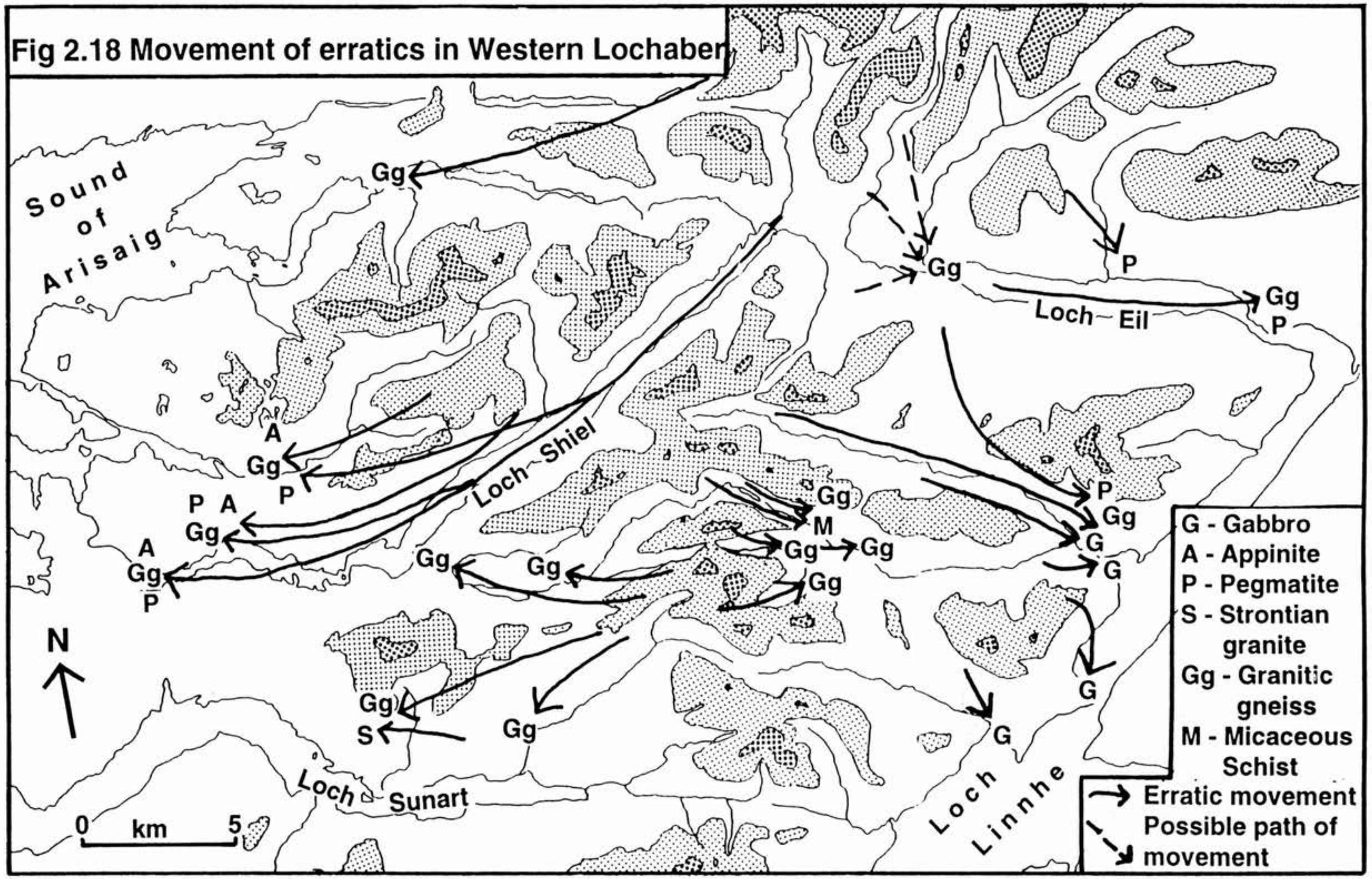


Fig 2.18 Movement of erratics in Western Lochaber

These results are consistent with ice movement down valleys, away from local mountain groups, towards the sea lochs. Ice flowed eastwards down Loch Eil, and both east from eastern Ardgour, and west from the mountains east of Loch Linnhe into this Loch. In southern Ardgour, the ice shed probably lay in the mountains to the east of the granitic gneiss outcrop, whereas further north the ice shed was probably in the mountains on, or close to the outcrop.

2.3.6 Glacial Erosional Features

Striations and friction cracks

Potentially suitable outcrops encountered in the field on trimline and mapping traverses were examined briefly and the location and orientation of any markings recorded. Striations are most common in places where there is freshly exposed bedrock such as on the sides of estate tracks e.g. in Glen Finnan, along coasts where deposits have been eroded away e.g. Loch Shiel, or in quarries e.g. Acharacle, and at low altitudes. Striations are also found on outcrops close to present day sea or lake levels around the shores of lochs. In addition, there are a few striations elsewhere in the glens, these often being on polished quartz outcrops. Most of the striations are oriented down major troughs in the area, but there are also some indicating ice flow over high ridges.

The best examples of friction cracks are on the shores of Loch Ailort at Roshven (Fig 2.19). Here, crescentic gouges are found in close association with striations and glacial grooves on smoothed and polished roche moutonneé shaped outcrops, all indicating ice flow towards the mouth of the Loch (left of plate). Crescentic gouges are also present on slabs on the col east of Beinn Resipol indicating ice flow from north to south over this col.

Fig 2.19 Striations and friction cracks at Roshven, Loch Ailort

Roche Moutonnées

The distribution of roche moutonneé is shown in Fig 2.1. Those found were of 2-15m long axis, and up to 3m high. One, alongside Loch Sunart (NM 829 609), has now been entirely removed by a recent road widening scheme. They are absent from the north eastern glens and are mainly found on the floor and lower valley sides of the western halves of the western glens, and on the crest of a low mountain ridge and summit in the west of the area. These all indicate westerly flowing ice. In contrast, bedrock mounds on the base of the eastern glens have no steep plucked face (e.g. NM 935 725, NM 970 634). In addition, there are numerous large bedrock outcrops on the glen sides which have clear plucked and abraded faces. These are found in the Eilt trough, Glen Finnan, Gleann Dubh Lighe, and on the col to the north of Glen Fionnlaighe. In all these cases the features have abraded and plucked faces which indicate ice flow down the troughs they occupy.

Smoothed slabs

Slopes on which a large proportion of the ground was covered by large, bare smoothed slabs were mapped and the distribution of these is shown in Fig 2.1. These slabs are of at least $10m^2$ and have fairly even, planar surfaces (see plate A in Fig 3.5). Smoothed slabs are widespread on the sides of glens in the field area, particularly in narrow, steep sided glens, in the west and south, and along the main mountain axes. In particular, the walls of Glens Gour and Tarbert, Coire an Lubhair, the upper Shiel trough, and much of the Eilt/Ailort trough consist almost entirely of smoothed slabs. Smoothed slabs are also common on spurs that project into valleys, for example in the Strontian/Gour trough, and the Hurich/ Gleann an Lochain Duibh trough. Smoothed slabs are most common on lower valley slopes, but are found at all altitudes from below sea level to the highest summits. Smoothed slabs are less common in the eastern half of the former LLS ice cap than in Western Lochaber. This may be a function of burial beneath the thicker drift cover.

Meltwater channels

The large meltwater channels incised into bedrock in the study area are shown in Fig 2.1. These channels are 5-15m deep, 3-20m wide, of U-shaped cross section, 100-400m long, and of straight or slightly sinuous plan form. The have low surface gradients, and most are found in the lower sections of the troughs they occupy. They mostly dip seawards and cut across local bedrock ridges which obstruct valley floors, or occasionally cross interfluve ridges, linking adjacent troughs.

In addition to these channels, there are numerous small channels cutting across many of the summit ridges in the study area. These are smaller, typically 3-5m wide, 15-200m long, and 1-10m deep (Fig 2.20). The bases of these channels are often covered with small clasts. The directions of dip of the base of these channels do not always show consistent patterns; adjacent channels on several spurs slope in different directions, so the interpretation of palaeoflow directions must be open to question. These diametrically opposite slope directions are glaciologically feasible because it is possible for sub-glacial meltwaters under high pressures to flow uphill. Comparison with geological information available for the study area suggests that many of these smaller meltwater channels have exploited geological faults. The morphology of these features is entirely consistent with those described elsewhere (Sugden and John 1976), and a meltwater origin is argued here.

Fig 2.20 Small meltwater channel crossing summit ridge above East of northern Loch Shiel. Channel approximately 2m wide and 2m deep.

2.3.7 Raised marine evidence

The evidence for higher sea-levels in Western Lochaber is summarised in Table A.3.1 of Appendix 3, and the distribution of raised marine features is shown in Fig 2.21.The evidence is categorised according to implied mean sea-level altitudes at the times of formation of each feature. The data points are separated into four groups on the basis of age interpretations made according to the established palaeosea-level changes discussed in Section 1.4, and any additional evidence for the age or relative age of the measured features. It is assumed here that palaeosea-level changes in Loch Sunart were similar to those established by Shennan (1994) for the lochs to the north.

Raised marine features in the 'High' category are at an altitude which suggests formation prior to the LLS, ie. above ~8m OD in the Western sea lochs and above ~12.5m OD in Loch Linnhe. Those in the 'LLS' category are obtained from the altitude of inferred LLS glacigenic deposits which show evidence of subaerial deposition. Those in the 'Intermediate' category are around the altitudes attained during the Postglacial Marine Transgression (PGMT) (6.5 - 8m OD in the western sea lochs and ~10.5 - 12.5m OD in the east). Although MSL during the PGMT was at the same altitude as at some point during sea level fall prior to 10,000BP (see Fig 1.8), as there was more time available during the PGMT for clear beaches and storm bars to form, it is suggested that such features are likely to have formed during the latter period. The points in the 'Low' category are at an altitude or location (e.g. below local evidence in the 'Intermediate' category) which suggests they date from after the PGMT.

Fig 2.21 shows that 'High' raised beaches are restricted to the outer parts of lochs Ailort, Moidart and Sunart. This pattern is confirmed from additional evidence mapped by the BGS (Sheets 52 and 61, drift editions). The 'high' raised marine evidence on Shona Beag in Loch Moidart and at Resipole on Loch Sunart are both tentative and require discussion. The evidence on Shona Beag is a section (NM 668 739) at up to 9.33m above current MHWS showing mostly horizontally bedded sands, occasional gravels and a few larger clasts. The locational position of the section suggests the deposits might be part of a small tombolo and these sediments may be raised marine deposits with the clasts representing iceberg rafted debris. If this genetic interpretation is correct, they imply a paleaosea-level of >9.61m when deposited. Comparison with Shennan's curve suggests this was prior to ~11,000B.P. However, as a lacustrine, or subaerial environment of deposition cannot be ruled out, interpretation of the significance of these sands remains equivocal. At Resipole implied mean sea-levels of <11.2m and 9.3m O.D. are identified from raised marine evidence. If Shennan's

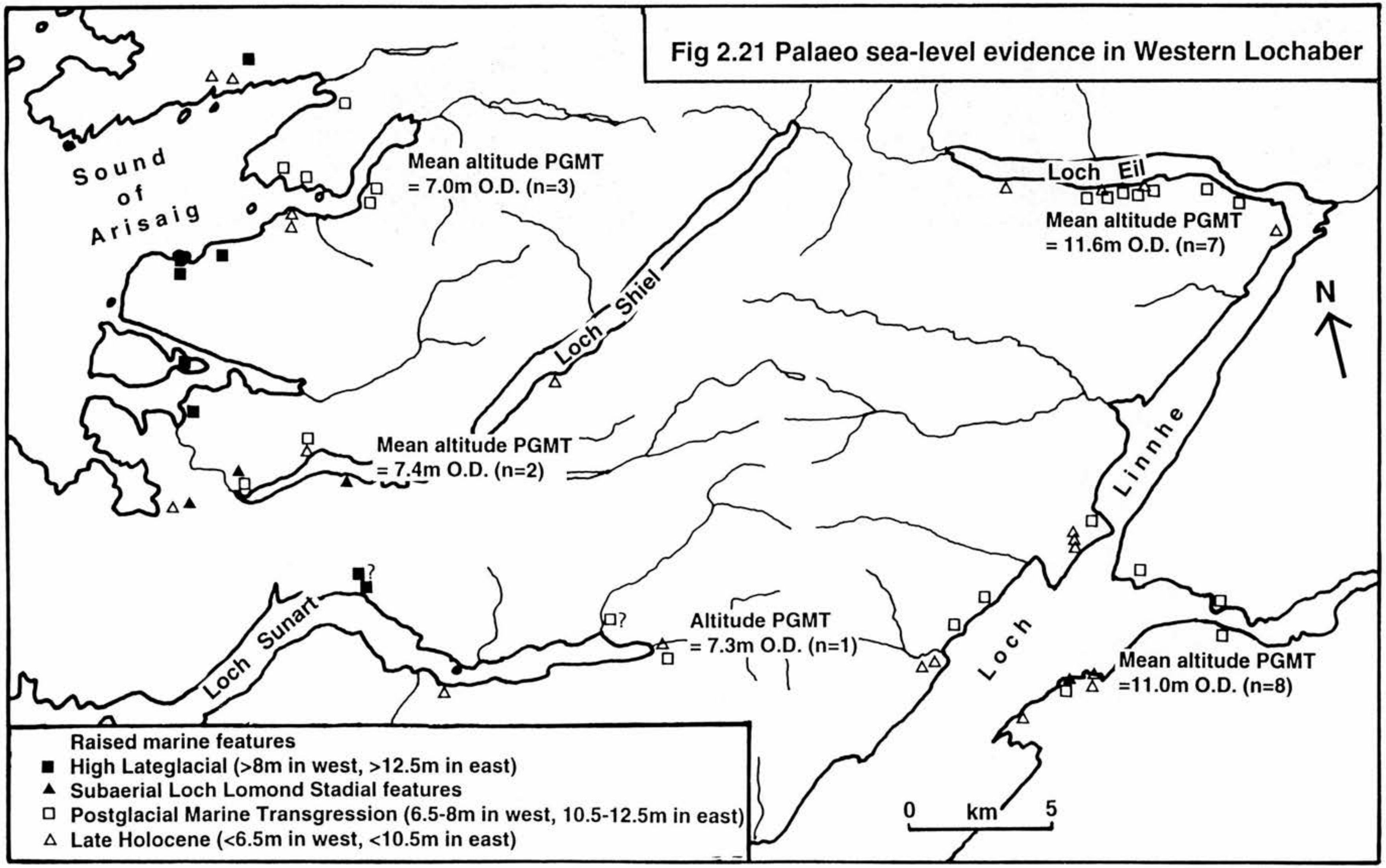


Fig 2.21 Palaeo sea-level evidence in Western Lochaber

sea-level curve applies to Loch Sunart also, these features may date from the early LLS or before, but as there is no sea-level curve for Loch Sunart, this interpretation must remain tentative.

Evidence of LLS palaeosea-levels is present in two locations. Several sedimentary sections (see Section 2.3.3) around the mouth of Loch Shiel show outwash gravels which are interpreted as being of fluvioglacial origin, deposited during the LLS. These occur at least as low as 6.63m O.D., which implies mean sea levels of below 4.5m O.D. at the time of formation, if the interpreted subaerial origin is correct. Near Kentallen on Loch Linnhe a beach ridge implying a former MSL of 11.2 m O.D. appears to grade into a moraine 1.8m lower than the bar which is located along the ice contact slope of an outwash fan (Fig 2.13). The continuity of the beach ridge and moraine forms could be interpreted as indicating the synchroneity of their formation. This would imply that both LLS and PGMT sea levels were around this altitude.

Mean values for implied mean sea-level during the Postglacial Marine Trangression (PGMT) from the different raised marine features in each sea loch are shown in Fig 2.21. There is widespread evidence of PGMT sea-levels in Lochs Linnhe and Eil, and all this evidence consists of depositional or erosional features which have affected LLS glacial deposits. There is rather less evidence in the western lochs, reflecting the general lack of glacigenic deposits to be reworked by marine action. The evidence includes well developed beach bars, for example that at Ballachulish, and erosional notches in till along the shores of Loch Eil, and notches in outwash such as at Corran. These features suggest that PGMT sea-level attained ~7.2m OD in the Western sea lochs, ~11.0m OD in Loch Linnhe and ~11.6m OD in Loch Eil, and thus supports existing models of sea-level change (Section 1.4). These differences between the PGMT altitudes in between the east and west of Western Lochaber may reflect differential rates of postglacial isostatic rebound, as they are broadly consistent with current ideas as to the spatial patterns of postglacial rebound (Lambeck 1993b, Firth et al. 1993, Sissons 1983).

In several locations there is additional sea-level evidence below features attributed to the PGMT, which indicate MSLs during subsequent regression. Many of these features are also superimposed on LLS glacial deposits, and others are concentrated around the Sunart narrows (see also BGS sheet 61, drift edition).

In addition to the evidence shown on Fig 2.21, it is probable that the flat areas at the heads of lochs Eil, Ailort, Sunart, Shiel and Moidart, and the mouth of Glen Gour, are raised estuarine deposits. River bank sections at the heads of Lochs Eil, Shiel and Moidart and in Glen Gour, show silty and sandy beds with few clasts which are typical of estuarine sediments. These flat areas all lie below ~10m O.D., which is consistent with formation during the PGMT.

The main rock platform is a prominent coastal feature south of the study area and less well developed stretches and fragments of the platform have previously been identified in Western Lochaber. In this study, mapping confirmed that the raised cliffline is a near continuous feature on both sides of upper Loch Linnhe (Peacock 1975, c.f. Sissons 1975), and that it also occurs in smaller fragments on the shores of Lochs Moidart and Ailort (Dawson 1988), and Loch Sunart (Fig 2.1).

2.4 Discussion

2.4.1 Reliability of the evidence

Due to repeated air photograph and field mapping, it is likely that the landforms mapped represent a fairly comprehensive and accurate representation of the glacial geomorphology of the area. There are four areas in which high levels of completeness or accuracy may not have been achieved.

Firstly, hummocky moraine may not have been accurately interpreted as being of a chaotic or aligned spatial arrangement. It was found particularly difficult to map hummocky moraines from air photographs, and to distinguish depositional hummocks from ones cored by bedrock. There are many of the latter in the field area. It is also not easy to separate dissected drift from low relief moraine hummocks; hummocks are present in a continuum of forms from dissected drift to constructional morainic ridges. It is likely that the main areas of hummocky moraine have been correctly and comprehensively mapped, but it is possible that some hummocks are bedrock cored, and that some marked as chaotically arranged may be more properly described as aligned.

Secondly, detailed sedimentary evidence was not collected; more intensive investigations are likely to uncover additional information with which to augment and refine the interpretations given.

Thirdly, striations have not been comprehensively mapped.

Fourthly, there are likely to be errors in the reconstructed palaeosea-levels, for several reasons. There will be some inaccuracies in measuring the altitudes of raised marine features due mainly to inaccuracies in locating MHWS from seaweed evidence, partly as local MHWS varies temporally (Kidson 1982). In two instances the altitude of the most prominent high seaweed line, interpreted as the MHWS altitude, was measured using the E. D. M. from O.S. benchmarks. These revealed differences of +0.18m and +0.63m compared with the admiralty chart MHWS altitudes at Ballachulish and western Loch Eil respectively. The altitudes measured by Electronic Distance Measurer are thus used wherever possible. Error is also introduced in inferring contemporary MSL from the nature of the raised marine evidence.

Relationships between various morphological features and MSL vary according to meteorological conditions, the results of unknown low frequency, high magnitude storm events and the degree of exposure of the coastline (Kidson 1982, van de Plassche 1986). There are also errors involved in estimating MSL and MHWS at locations remote from the primary and secondary ports for which data is published by the Admiralty. In addition, palaeotidal ranges may have been different under conditions of different sea-levels (van de Plassche 1986).
As a result of all these possible sources of error, it is estimated that most of the reconstructed palaeo-MSL altitudes presented in Appendix 3 and Fig 2.21 are likely to be accurate to within +/- 1.5m, depending on the location, type of evidence and the method of measurement.

2.4.2 Summary

There is very little glacial drift and few moraines in Western Lochaber, compared to the situation further east. Glacial drift is most widespread in the northeast of the area mapped here, and elsewhere is only found on the lower slopes of the glens, where sedimentological evidence suggests resedimentation by downslope sediment gravity flows. In the west and south of Western Lochaber, particularly along the shores of the lochs, drift deposits are very sparse. Few maximum limits in Western Lochaber are marked by terminal and/ or lateral moraines. The largest volumes of glacial deposits are in the form of proglacial outwash fans. These are found in the shallow parts of the sea loch troughs. The largest fans are found in the Linnhe and Shiel troughs, which have the largest catchments at present. Signs of glacial erosion in Western Lochaber are particularly widespread, and occur at all altitudes. Raised marine features indicating palaeosea-levels of <13m OD in the east, and <8m in the west are found in association with glacial deposits in the inner sea lochs, whereas higher raised beach evidence is confined to the outer lochs.

Chapter 3 - Trimline evidence

3.1 Aim

The aim of this chapter is to map trimlines in order to reconstruct the surface morphology of the LLS ice cap in Western Lochaber. First, it is necessary to establish that trimlines exist and can be used systematically.

3.2 Background

Glacial trimlines record former upper ice margins of valley glaciers or ice caps in areas with numerous nunataks. There are two types of trimlines, one involving contrasts in the degree of vegetation development, and the other contrasts in the amount of weathering and frost shattering of bedrock surfaces. Both are useful tools for reconstructing the thickness of former ice masses.

Vegetation trimlines are found in glacial regions where ice flows into heavily vegetated areas, and the glacier has thinned or retreated from its maximum extent leaving an abrupt change in the vegetation cover above and below the position of the former glacier surface. Such trimlines are found in association with contemporary glaciers in Patagonia (Mercer 1965), the Colorado Front Range (Meierding 1982), Baffin Island (Hawkins 1985), New Zealand (Chinn 1975) and the European Alps (Grove 1988).

A bedrock trimline separates ice smoothed bedrock below from weathered and frost shattered outcrops above it. This reflects the difference between glacial erosion below and subaerial periglacial action above the line. They are useful for reconstructing ice limits in the upper reaches of former glaciers, where there are no lateral moraines of drift limits. Bedrock trimlines can be preserved as relict features if the site has escaped subsequent subaerial modification, and in some areas bedrock trimlines have been used to reconstruct the dimensions of former glaciers, for example in Tasmania (MacIntosh, unpublished), Norway (Nesje et al. 1987), and Eastern Arctic Canada (Ives 1975). In some instances the trimline may be marked by clear contrasts in both vegetation development and bedrock surfaces, as is the case on Cerro Pearson in Argentinean Patagonia, where the vegetation trimline was noted by Mercer (1965).

In Scotland, trimlines have been discussed and used most widely by Thorp (1981, 1984, 1986). He mapped the distribution of glacial and periglacial features on traverses up the extremities of spurs, and suggested that the distinct altitudinal limit separating them recorded the LLS maximum ice surface (Fig 3.1). The values obtained from mapping the evidence on spurs was supplemented by recording the glacial or periglacial evidence present on nearby low mountain summits and cols. By mapping trimlines in this manner, he reconstructed the surface of an ice field in an area immediately to the east of Ardgour and Moidart. Trimlines have also been used to map the vertical extent of former ice masses in Scotland by Ballantyne (1989) on Skye, Sissons (1977) and Reed (1988) in the northern Highlands and Tate (unpublished) in the north west Highlands (Ballantyne, pers. comm.).

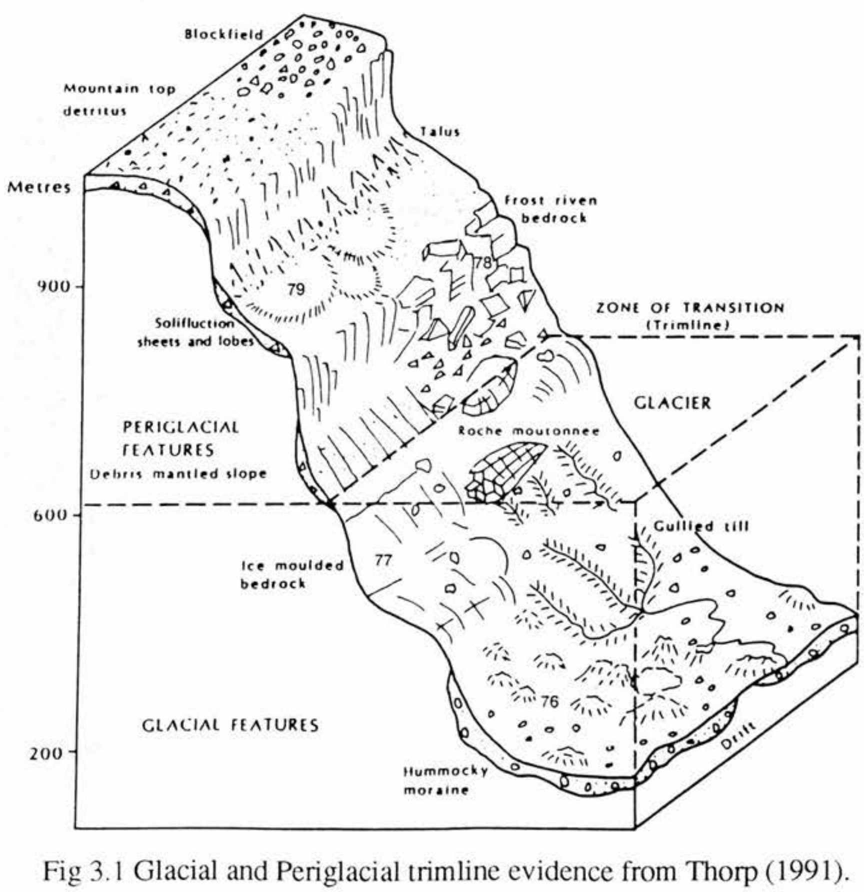


Fig 3.1 Glacial and Periglacial trimline evidence from Thorp (1991).

There has been some controversy over interpretations of bedrock trimline evidence. Workers such as Nesje et al. (1987, 1988) working in Southern Scandinavia, and Pheasant and Andrews (1972) working in Baffin Island, have argued that trimlines represent the contrast between lower slopes that have been under ice and subject to glacial erosion and deposition, and upper slopes that were above the glacier surface and were subject to severe periglacial action. Sugden and Watts (1977), however, argue that the weathering contrast could reflect the difference in basal thermal regimes beneath a former ice sheet. They propose that on the plateaux glacial erosion was ineffective under thin cold based ice. As a result pre-existing weathered and frost shattered debris survived intact beneath cold based ice, allowing landforms dating from before the last glaciation to be preserved. Thicker, warm-based ice in the valleys, by contrast, produced widespread glacial erosion. Thus contrasts between glacial scouring in the valleys and weathering on higher plateaux may not necessarily represent differences associated with former glacier margins. Kleman (1994) similarly argues that the distribution of cold based ice under a former ice sheet can explain the weathering zones in southern Scandinavia in a more satisfactory manner than the former maximum ice surface altitude explanation proposed by Nesje et al. (1987). This cold-based ice hypothesis suggests that the former presence of cold based ice in certain locations may mean that glacial erosion is inefficient and preglacial landforms may be preserved. Since basal ice temperatures are largely influenced by ice thickness and ice velocity, cold based ice would be expected where ice is thin and where ice is divergent, and flowing at lower than average velocities. For these reasons cold based ice will occur preferentially over hills and massifs, as shown by Glasser (1992) in the Eastern Grampians. Under such circumstances, one would expect relict regolith to be preferentially preserved on high hills. In addition, the relict weathering will be preserved at lower altitudes in zones of divergent ice flow than in areas of convergent ice flow.

There are several possible means of evaluating the status of the trimlines in Western Lochaber. One approach would be to examine the glacial evidence for former basal thermal regimes during the Devensian and Loch Lomond Stadial glaciations. In the maritime environment of western Lochaber it is unlikely that there was cold based ice. Rather, there is widespread evidence that even the mountain summits were covered by actively eroding ice during the Devensian glaciation. This evidence is in the form of numerous relict smoothed slabs, ice moulded outcrops and roche moutonnées, occasional striations, and many meltwater channels criss-crossing ridges, including the summit ridges of the mountains (Section 2.3.6). It thus seems reasonable to assume that the Devensian ice sheet was warm based throughout Ardgour and Moidart. If a similar or smaller sized ice cap covered the area during the LLS, it is again reasonable to assume that it too would be warm based throughout. This is supported by widespread evidence of fresh glacial erosion, such as unweathered smoothed slabs, striations and roche moutonnées at lower altitudes.

Secondly, if the spatial distribution of the trimlines was a function of basal thermal regime, then the altitudes would show an association with topography. In areas where ice flow was divergent, such as at the ice sheds in the mountains, ice would tend to be cold-based, and trimline values would be lower than where ice flow was convergent, and warm based. By contrast, if the trimline values obtained represent a former ice surface, the distribution should reflect former ice surface gradients. An examination of the trimline values presented later in this chapter shows that the values are highest along the main mountain axes and lowest at the seawards ends of the glens. This pattern is again consistent with the interpretation that they represent the altitude of a former ice surface, and does not support the alternative that they are related to the basal thermal regime of a former ice mass.

Thirdly, concentrations of glacially transported debris are often found at the trimline altitude, and these cannot be explained if trimlines were a result of contrasts in basal thermal regime.

A second possible means of investigating the origin of the trimlines would be to date rock surfaces above and below the trimlines using either O.S.L. dating or cosmogenic ^{10}Be surface exposure dating of quartz grain techniques. The latter means has been used to corroborate trimline interpretations in Norway by McCarrol and Nesje (1993). This has not been attempted here, although these techniques are now possible.

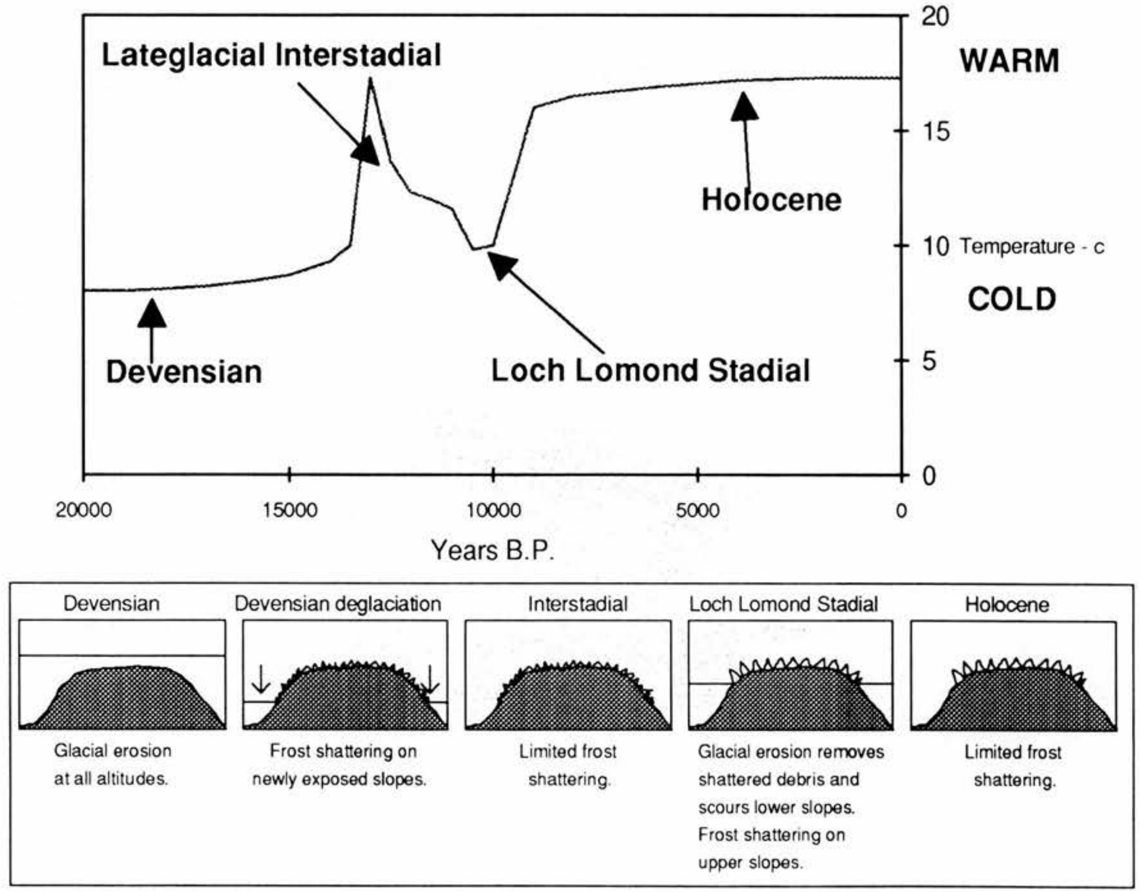


Fig 3.2 Evolution of LLS trimlines. Lateglacial temperature curve from Coope (1977), and schematic illustration of episodes of ice cover and frost shattering.

Assuming that any trimlines relate to the LLS glaciation, then one can use a knowledge of late-glacial climate and ice volume changes to suggest some morphological relationships (Fig 3.2). Evidence already discussed suggests that this area was completely covered by active ice during the Devensian glaciation, resulting in a landscape of areal scouring. As deglaciation proceeded at the end of this glacial period, some valley slopes were probably exposed to frost action until full interstadial climatic conditions were attained. When the climate deteriorated once more and glaciers advanced during the Loch Lomond Stadial, any existing periglacial debris would have been entrained by advancing glaciers. The bedrock would have been smoothed and scoured by ice erosion, and moraines and drift deposited. On exposed slopes above the glacial limits the severe climatic conditions of the stadial would have resulted in further intense periglacial activity. Climatic amelioration following the LLS was rapid (Coope 1977, Dansgaard et al. 1989), and so during deglaciation there was probably only limited periglacial activity on slopes that were progressively exposed as the glaciers retreated. Evidence suggests that under contemporary climatic conditions, which are typical of those which have prevailed throughout the Holocene, periglacial activity is restricted to small scale processes and severe large scale frost shattering or lobe formation does not occur (Ballantyne 1984). The Loch Lomond Stadial trimlines would therefore have been preserved as relict features.

not above ice

3.3 Trimline morphology and clarity

The morphology and clarity of the trimlines depends on the nature and intensity of the periglacial and glacial processes that prevailed during the Loch Lomond Stadial on the particular hillslope, and on any subsequent downslope movement of debris. This section assesses the effects of variations in the intensity of periglacial and glacial processes on trimline morphology and the effects of topography and geology on their development and preservation. The result is a series of predictions as to locations where trimlines are likely to have been best developed and preserved, and hence which valley slopes are likely to show the clearest evidence.

It is likely that the intensity of periglacial processes varied across the ice free parts of Western Lochaber during the LLS. In a relatively warm maritime climate the severity of periglacial action increases with altitude, as temperatures correspondingly decrease. The Environmental Lapse Rate (ELR) under contemporary climatic conditions is 0.6^{o} C per 100m (Harding 1978). Climatic reconstructions for Scotland during the LLS suggest that mean annual air temperatures at sea-level may have been approximately -8^{o}C to -10^{o}C (Ballantyne 1984). This suggests mean annual air temperatures at 600m and 800m of -11.6^{o}C and -12.8^{o} C respectively, assuming the same ELR. Estimates of mean July sea-level temperatures range from 5^{o}C (Sissons 1976) to 8 or 9^{o}C (Coope 1977). The corresponding temperature at 800m would be $\sim$+1^{o}C. According to

Hallet (1983) frost cracking of bedrock can occur at temperatures of -5 to -15^{o}C, providing there is sufficient rock moisture. The climate for most of the LLS was maritime (Ballantyne and Harris 1994, Kerr 1992), and so it is reasonable to assume that moisture was abundant. Mean annual air temperatures of -6 to -8^{o}C are necessary for the development of continuous permafrost, and -2 to -8^{o}C for discontinuous permafrost (Ballantyne and Harris 1994). It is thus likely that the LLS climate was sufficiently severe for the widespread development of permafrost at all altitudes, and for seasonal frost cracking to occur. As a result, one would expect to find lobes and terraces, deep bedrock joints, shattered outcrops, and scree.

The type and clarity of glacial evidence left by a former glacier will reflect the glacial regime, and vary longitudinally down the former glacier long profile. These changing glacial processes will affect the morphology and clarity of the trimlines. The efficiency of debris entrainment and bedrock erosion at the ice margin will depend partly on the length of time the glacier maintained its maximum position and on the rate and duration of ice marginal fluctuations. These factors will influence of the width of the trimline zone. Ice sheet modelling experiments by Payne (1988) suggested that maximum volumes for the LLS ice cap were sustained fleetingly before being truncated by rapid volume loss as the climate warmed suddenly. If correct, the maximum ice extent may have been maintained for only 100-300 years which may not be long enough to produce widespread glacial evidence at the maximum ice surface altitude. Instead, there might be a zone of minor glacial modification at and immediately below the former maximum ice surface altitude. Therefore evidence of well scoured rocks and removal of loose shattered debris may underestimate the true trimline values.

The intensity of various glacial processes also varies longitudinally down the former glacier long profile. This means that the character of the trimline is likely to vary downglacier. Table 3.1 summarises the theoretical relationships. It is based on concepts of the changing longitudinal ice flux and whether or not flow at the margin is towards the margin (compressive) or away from the margin (extending). Compressive flow results in deposition at the margin, and this predominates below the ELA of land terminating glaciers. In these areas the extent of drift deposits and lateral moraines may be used to reconstruct the dimensions of former glaciers, rather than trimline evidence. Extending flow is associated with debris entrainment and erosion, and predominates in the accumulation areas of land terminating glaciers, and throughout the length of calving glacier margins (Powell 1991). The ice flux at any point is important as this controls the volume of debris transported past the point as a potential agent of abrasion. In a land terminating glacier ice flux increases to a maximum at the ELA, and decreases to zero at the snout. In a calving glacier, the ice flux may increase continuously from the ice shed to the snout.

Table 3.1 Likely conditions at the upper surface margins of an idealised glacier.

Section of glacier	Ice velocity	Ice flow	Debris entrainment	Erosion	Deposition
Ice shed	zero		zero	zero	zero
Accumulation area	increasing	extending	yes	yes	zero
Around ELA	high	uniform	yes	some	some
Ablation area (land terminus)	decreasing	compressive	little	zero	yes
Ablation area (calving terminus)	high	extending	yes	yes	some

At the ice shed the ice velocity is zero; hence there is very little debris entrainment and no glacial erosion. Glacial evidence is unlikely to be clear here. In the accumulation area ice velocities increase towards the E.L.A. and flow is extending. There will therefore be some entrainment and transportation of debris and some erosion, although not necessarily up the level of the maximum ice surface, and these processes will become more efficient down glacier. Around the ELA there are high ice velocities and ice flow is neither extending nor compressive, so both erosion and deposition may occur. If the glacier has a land based terminus, ice velocities decrease from the ELA towards the terminus, and ice flow is compressive, so deposition is widespread around the ice margins. However, where the glacier has a calving terminus ice velocities remain high and flow may remain extending right to the terminus; the glacier is a relatively efficient transporter and eroder of debris. Deposition is concentrated at the calving front , with less at the lateral ice margins.

During deglaciation a glacier recedes and the ELA moves back towards the accumulation area. This means that depositional ice marginal features are superimposed upon the landscape at successively lower altitudes, and the slopes below the former ice surface maximum become progressively exposed to glacial deposition.

These relationships may be summarised as follows. At the ice shed trimlines are likely to be poorly developed and only defined by lower limits of extensive periglacial evidence, although previously shattered blocks and debris will not necessarily have been removed from below the line.

Downglacier, but still in the accumulation area, there will be increasing signs of glacial erosion at and below the trimline and most pre-existing periglacially generated debris will have

been removed. The trimline is likely to be visible as a zone in which the amount of glacial modification decreases upwards. Glacial deposits will only be found well below the trimline altitude as a result of deposition during deglaciation.
Around the ELA debris should have been efficiently removed, and signs of glacial erosion widespread. Glacial deposits may be found as high up the slope as the trimline.
In the ablation area of a land terminating glacier there will not be erosional trimlines, but the former ice surface altitude should be marked by a zone of glacially deposited debris, with glacial deposits widespread below the zone.
In the ablation area of a calving glacier there may be extensive signs of glacial erosion and possibly a zone of glacially transported debris at the trimline.

Topography can also influence the clarity of trimline evidence. Thorp (1981) noted that the best trimlines are found on spurs projecting into troughs. He argued that this is partly as ice flow is confined in such places, resulting in a higher ice flux, and enhanced potential for ice scour. Glacial erosional evidence below the trimline is thus likely to be clear in such locations. As his field area contains numerous high, steep sided spurs he only utilised spurs for trimline evidence.
In this study both spurs and slopes in narrow, straight valley sections were used, providing the slope angle remained fairly constant throughout. This avoids the problem of complications due to changing topography.

Topography will also influence the preservation of glacial and periglacial evidence. Where slopes are steep, it is probable that periglacially derived debris and glacial deposits will be moved downslope subsequent to their formation/deposition. Thus on steeper slopes the minimum altitude of periglacial debris and the maximum altitude of glacial debris may both be below the actual former glacial maximum.

Geological variations will influence the clarity of trimlines. For trimlines to be clear, the bedrock must be sufficiently weak for periglacial action to produce weathered and shattered outcrops and glacial action to mould and scour outcrops during the LLS, yet sufficiently resistant that weathering and frost action during the Holocene will not have obscured the differences formed during the LLS. Lithology and structure both exert a tremendous influence on the degree of weathering that takes place on a bedrock outcrop (Table 3.2). Some rock types become frost shattered under contemporary climatic conditions, for example fissile semipelitic and pelitic micaceous schists (Hallet 1983), whereas others, for example massive gneisses, are resistant to frost action, even under the severe climate prevailing over mountain tops during the Loch Lomond Stadial. Neither of these rock types would produce clear trimline evidence; nor would a slope where the bedrock varies in resistance to weathering up the slope. The clearest trimlines are found on slopes where the rock type is uniform throughout, and where this rock is neither extremely susceptible nor extremely resistant to frost action. Many of the Moinian schists and gneisses which underlie most of the study area are ideal in this respect.

Table 3.2 Qualitative ranking of rock types in Lochaber according to susceptibility to frost shattering, after Thorp (1984).

Highly resistant	**Strongly resistant**	**Moderately resistant**	**Weakly resistant**
←	Rannoch moor granite	→	
		←	Fault intrusion granite →
←	Andesites, Rhyolites		→
		← Quartzites	→
←		Schists	→
	←	Psammites	→
Large ←	**Size of frost riven blocks**		**→ Small**

3.4 Methods

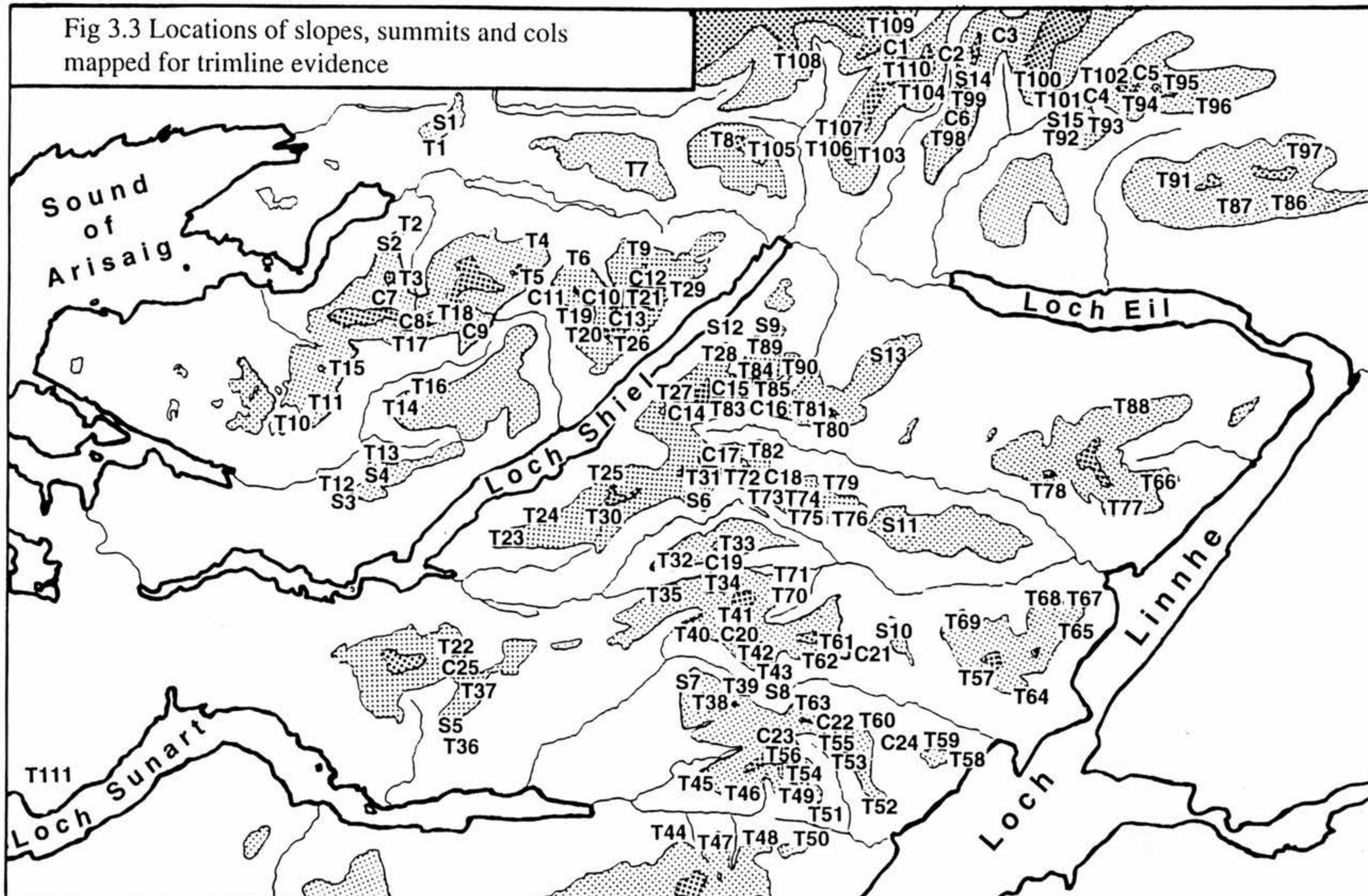
Slopes and spurs were selected for mapping on the basis of the criteria above, and the need to obtain an even distribution of trimlines for the purpose of reconstructing the former ice surface. 111 slopes in Western Lochaber were selected for examination, including one that was thought to be well outside the former ice limits as a methodological test (Fig 3.3). A checklist (Table 3.3) was used to note the presence or absence of various periglacial and glacial features at different altitudes as each slope was ascended. This checklist was based on that of Thorp (1981), with alterations following preliminary mapping to incorporate those features which were most useful and widespread in Western Lochaber. The glacial features recorded include drift and moraine deposits, accumulations of perched blocks or blocky debris (unrelated to steep rock faces or breaks in slope), smoothed unweathered slabs and striations. Periglacial features noted include deep and wide open joints, small boulder and solifluction sheets, vegetated fossil scree, flaggy and/or in situ frost shattered debris and weathered slabs. The development of several of these bedrock features are dependent on lithology and hence changes in abundance of these features on slopes of uniform geology are of more significance than comparisons of their development between slopes. The periglacial and glacial evidence on low summits and cols was also recorded, as advocated by Thorp (1986), in order to corroborate trimline evidence

Table 3.3 Trimline Checklist

Hillslope	*Seann Chruach, NE slope*			
Altitude	*50m*	*150m*	*200m*	*220m*
Drift mounds or sheets				
Hummocky moraines				
Roche moutonnées				
Perched blocks				
Blocky debris present				
Concentrations of debris not at breaks in slope or under cliffs				
Ice moulded bedrock - unweathered				
- weathered				
Smoothed slabs - unweathered				
- weathered				
Weathered and irregular bedrock surface				
Extensive areas jagged frost-riven bedrock				
Frost widened joints				
Rounded edges to blocks				
In situ frost shattered/weathered blocks				
Large angular frost riven blocks at base of outcrops				
Boulders in fossil solifluction lobes, terraces and sheets				
Ground mantled with platy/flaggy debris				
Thick angular scree				
Vegetated or lichen covered scree				
Stone stripes or circles				
Platy/flaggy debris present				
Position on slope				
Bedrock jointing				
Bedrock type				

In order to derive a quantitative test of trimline altitudes, two approaches were used. Firstly, measurements of joint depths were made from bedrock outcrops on most slopes where rock exposure and conditions allowed. Twenty joint depths were recorded from outcrops below, around and above the inferred trimline. Thorp (1981, 1984) and Ballantyne (1982) have suggested that the depths of open joints in bedrock outcrops are significantly higher above trimlines than below them. Ballantyne found that on sandstones at An Teallach joints more than 15-20cm in depth were only found above the trimline. This pattern only held above ~500m metres, since below this altitude deep joints were not found even outside the former glacier limits. Harrison and Edie (1992 unpublished) produced joint depth measurements from a variety of lithologies suggesting that the relationship with trimlines was consistent, with the actual joint depths being strongly influenced by bedrock type.

Fig 3.3 Locations of slopes, summits and cols mapped for trimline evidence



Secondly, an alternative quantitative approach was developed to distinguish between slabs that are above or below the former trimline. There is usually a clear difference between slabs that are weathered and those that remain fresh and smooth (see Fig 3.5, Plates A and E). Most of the slabs on the higher summits are of the former type. The weathered slabs are mostly characterised by up to 3cm of relief along the foliation planes in the schists and gneisses. The two types of slabs can therefore be distinguished by measuring the relief on a planar section of the slab. This was done in the field by placing a 70cm rule flat on the slab, and measuring the depth from the base of the rule to the bedrock surface at 2cm intervals. The mean of the values thus obtained was calculated. This method is similar to the profile gauge measurements used by McCarrol and Nesje (1993) to contrast bedrock weathering above and below Norwegian trimlines. Slabs were selected to keep lithology, surface slope and aspect constant on each slope, as these factors may influence weathering rates.

In order to determine the trimline altitude, three sets of features were recorded on all slopes. These were the maximum limit of extensive glacial evidence, the minimum limit of extensive severely frost shattered rock, and the altitude of any particular concentrations of debris in between. These three altitudes allowed a best estimate of the former glacier surface, in the form of a specific altitude, or an altitudinal zone, depending on the clarity of the evidence. In accordance with the controls on trimline morphology discussed in the previous section, the evidence used varied downglacier. In the lower reaches of the former glaciers this best estimate was the altitude of a debris zone, or the upper limit of extensive glacial evidence. In the upper reaches of the glaciers the altitude of the lower limit of extensive periglacial evidence provided the best estimate.

The reliability of the evidence on each slope was classified as shown in Table 3.4 according to the sharpness of the upper and lower limits to glacial and periglacial evidence on a slope, and the width of the trimline zone.

Table 3.4 Trimline clarity

Clarity of contrast in distribution of glacial and periglacial evidence above and below trimline	Category
Very clear contrast in distributions of glacial and / or periglacial evidence over a zone ≤ 100m wide	Firm
Contrast in distributions of glacial and / or periglacial evidence over a zone ≤ 100m wide, **or** very clear contrast in glacial and / or periglacial evidence over a zone 100 - 200m wide	Probable
Contrast in distributions of glacial and / or periglacial evidence over a zone >100m wide, **or** very clear contrast in glacial and / or periglacial evidence over a zone > 200m wide	Indistinct

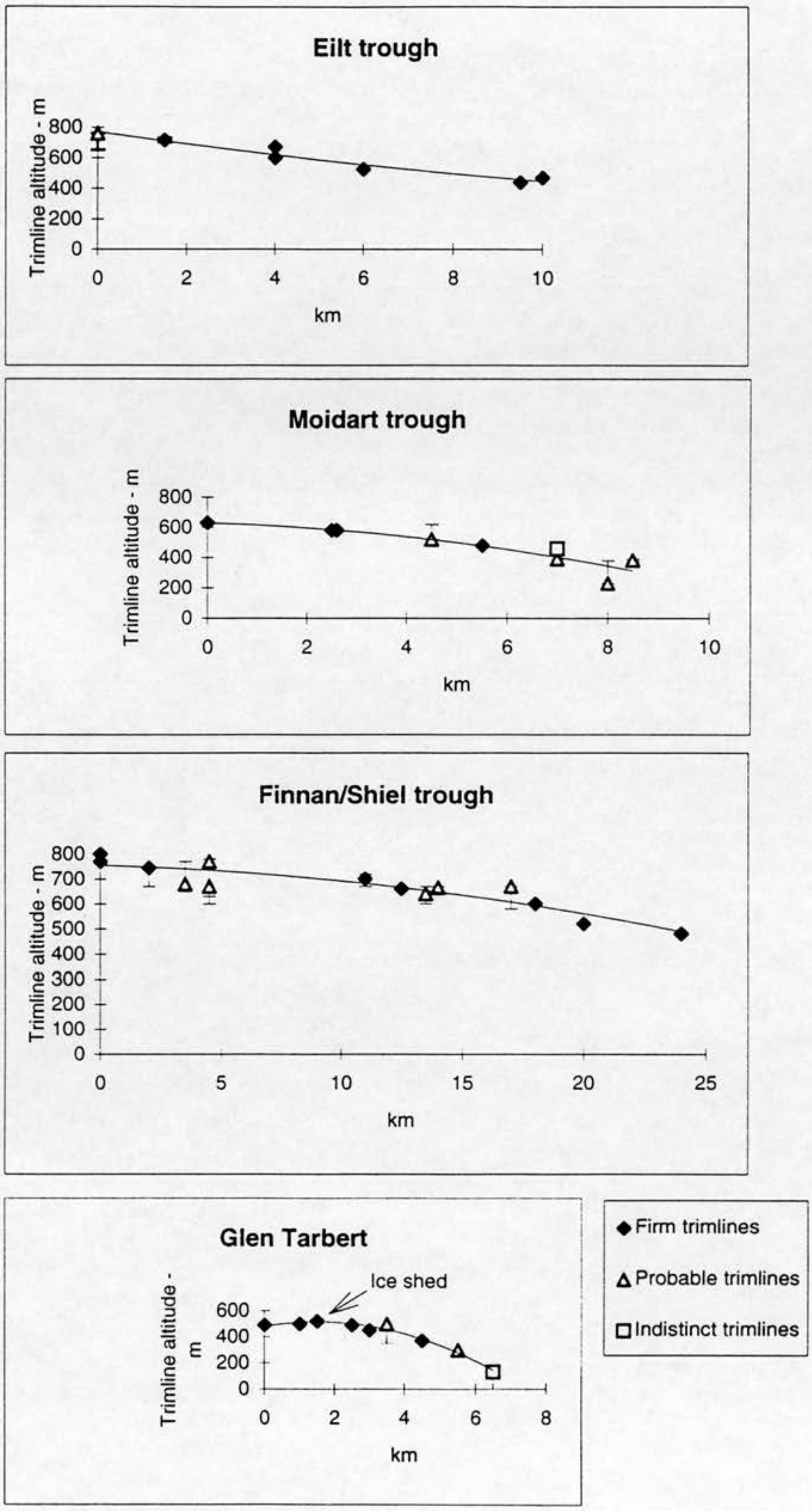


Fig 3.4 Trimlines altitudes and inferred LLS ice surfaces in glens.
Transects are approximately from watersheds to coasts.
Error bars indicate trimline zones where evidence is unclear.

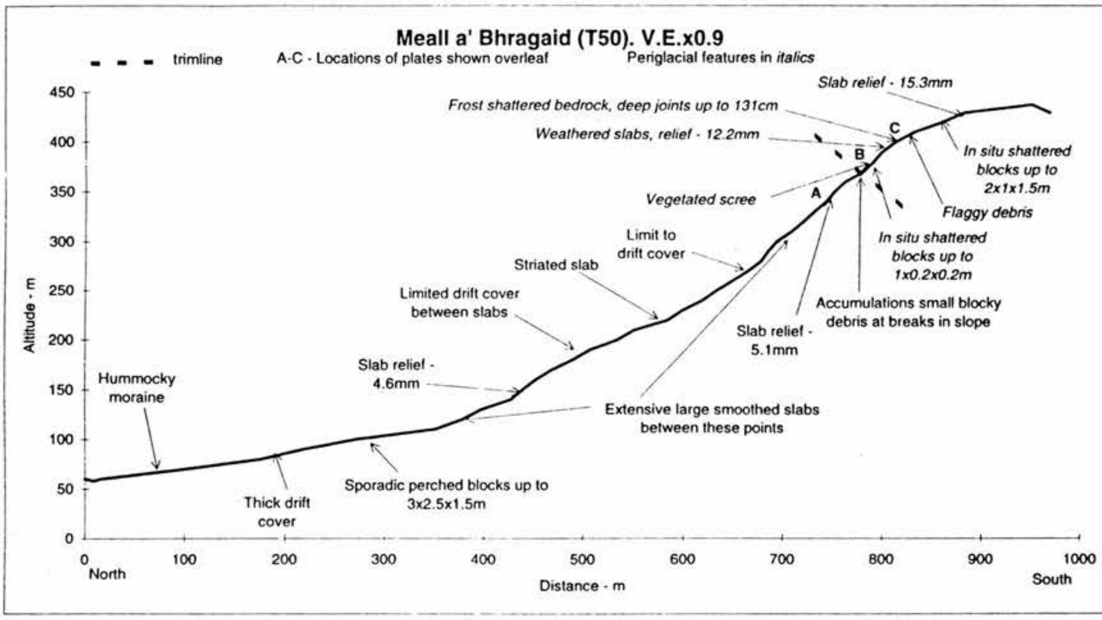
3.5 Results

3.5.1 Geomorphological features

The results of the trimline mapping exercise are summarised in Table A.4.1 in Appendix 4, and locations of the slopes, summits and cols surveyed are shown in Fig 3.3. On the majority of slopes there are clear contrasts in the distribution of glacial and periglacial evidence, so that fresh glacial features are only present below the trimline zone, and clear periglacial features are located only above it. In some instances the evidence suggests an accuracy of ± 5m, on other slopes the trimline zone is 150m wide, and elsewhere only a minimum or a maximum value for the trimline altitude can be obtained from the evidence on the slope. Of the 110 slopes ascended in the study area, 64 % show firm evidence, 25 % produce probable trimlines and on 11 % of the slopes the trimline evidence is indistinct. There is no evidence for a trimline on a slope in Ardnamurchan (T111) which lies outside the LLS ice limits; the distribution of glacial and periglacial features does not change with altitude. Fig 3.4 shows the locations of firm, probable and indistinct trimlines in some of the troughs (see also Fig 5.2).

It is helpful to discuss the reliability of the trimline evidence in more detail. Fig 3.5 shows the evidence found on two firm trimlines on which the trimline can be located to ± 5m. Each slope shows clear signs of fresh glacial action below the trimline such as moraines and unweathered scoured slabs (Fig 3.5, Plate A), and a zone of glacially transported debris 2-4m wide at the trimline altitude (Fig 3.5, Plate B). Above this there are numerous signs of small scale periglacial action such as deep joints (Fig 3.5, Plate F), in situ shattered blocks usually 0.5 - 1m^2 (Fig 3.5, Plate C), vegetated scree, small boulder sheets and lobes with risers of 0.5 - 1m and 5 - 10m slopes and thin, incipient mountain-top detritus consisting of angular flaggy debris. Relict glacial features above trimlines such as roche moutonnées and smoothed slabs have been weathered along the foliation planes (Fig 3.5, Plate E) and undergone frost action to produce deep joints (Fig 3.5, Plate F).

The firm trimline evidence found on the NE slope of Seann Chruach (T2), in the NW of the study area is an example. There are clear signs of fresh glacial action on the lower slopes. In the base of the valley are hummocky moraines and a thick drift cover which thins upwards to about 350m. The ground surface is largely vegetated but in places fragments of unweathered smoothed slabs are exposed. The surface is strewn with numerous large perched blocks of up to 3 x 1.5 x 2m in size (Fig 3.5, Plate D). There are particular concentrations of these at breaks in slope and they increase in abundance from 430 - 470m where there is a clear boulder line across part of the slope. This boulder line is not related to any crags above, nor is it at the base of a steep slope. A similar boulder line is present upvalley at 505m on the eastern flank of Coire a' Bhuiridh. These lines are interpreted as representing debris deposited at the margin



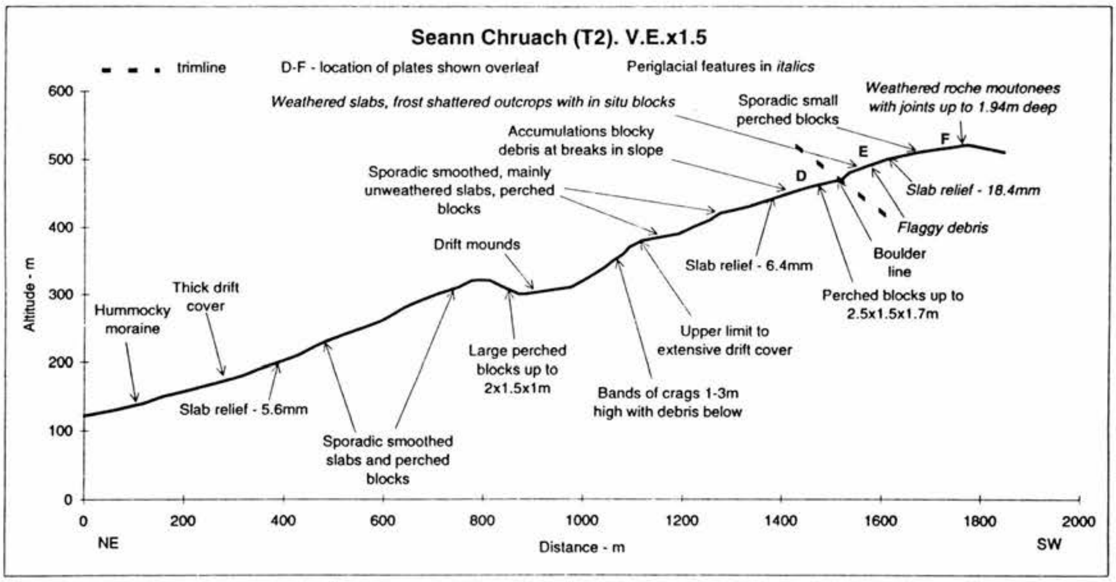


Fig 3.5 Transects up two slopes with very clear trimline evidence and plates showing glacial and periglacial features (following two pages).

Meall a'Bhragaid (T50) - Plate A. Large smoothed, unweathered slab. Ruler is 70cm long.

Plate B. Concentrations of debris at trimline zone.

Plate C. Frost shattered bedrock with in situ blocks and deep joints.

Seann Chruach (T2) - Plate D. Large perched blocks below trimline.

Plate E. Weathered slab, 70cm ruler indicates scale.

Plate F. Relict smoothed slabs and roche moutonnées split by deep, wide open joints.

A

B

C

D

E

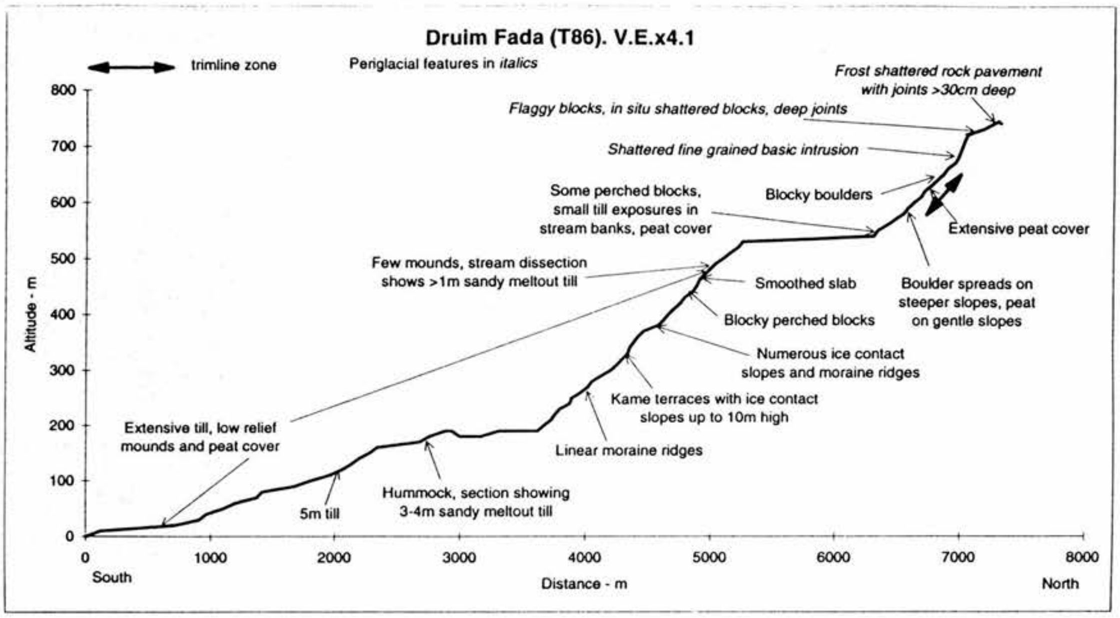
F

of the former glacier at its maximum. Above this boulder line widespread signs of periglacial action are found. From 470m to the summit at 520m there are numerous frost shattered outcrops, in situ blocks, large open joints of up to 1.94 m in depth, and numerous weathered slabs (Fig 3.5, Plate E). There are also weathered roche moutonnée forms on the summit, oriented towards the NNW and split open by joints of up to 1.94 x 0.5m (Fig 3.5, Plate F). The surface of these are littered sporadically with small perched blocks and quartzite and mica schist erratics.

Fig 3.6 shows the evidence on two slopes with wider trimline zones, and illustrates some of the reasons why trimline evidence may be less clear. On both slopes there is fresh glacial evidence below the trimline zone, and periglacial evidence above the zone. On Druim Fada there are glacial deposits up to 580m, and clear periglacial features above 650m, but the slopes between are mantled with thick peat and vegetation obscuring any evidence on underlying bedrock outcrops, so that the trimline altitude cannot be defined precisely. On Beinn an Tuim there are clear contrasts in the presence of glacial and periglacial evidence above and below a trimline zone 180m wide. It is likely that the evidence does not allow more precise reconstruction of the trimline altitude for two reasons. Steep slopes have resulted in widespread downslope movement of debris, so that glacially transported and frost shattered debris may no longer be in situ. Secondly, most of the outcrops in this zone are free faces which are usually frost shattered at all altitudes in Western Lochaber, and thus do not show clear trimline evidence.

Fig 3.4 shows that both where there are wide trimline zones, and where the trimline evidence is not firm, the altitudes still conform with the ice surface altitudes suggested by firm trimlines (see also Fig 5.2). Many of the less clear trimlines are located on slopes which were examined in order to obtain an even coverage of trimlines, but where there may be few bedrock exposures, thick peat cover, unusually resistant or unresistant rock types, changes in slope angle, or on very steep or very gentle slopes. These are all situations in which trimline evidence is unlikely to be clear or widespread.

Both the clear and the probable trimline altitudes cross-check with those on adjacent slopes and those across glens, giving confidence in the interpretations. Where trimlines were obtained from slopes on opposite sides of glens these cross-check to within 50m, for example T22 and T37 between the Shiel and Sunart troughs are at altitudes of 480m and 460m respectively. Furthermore the spatial pattern shows consistency in that values obtained from adjacent slopes decrease towards the sea lochs in the west and south, as shown in Fig 3.4. The altitudes of firm and probable trimlines in each of the glens illustrated both decrease down-valley and indicate plausible ice surface gradients. There are three locations where the evidence suggests anomalous trimline altitudes, namely the mountain massifs to the SW of Glenfinnan, the west



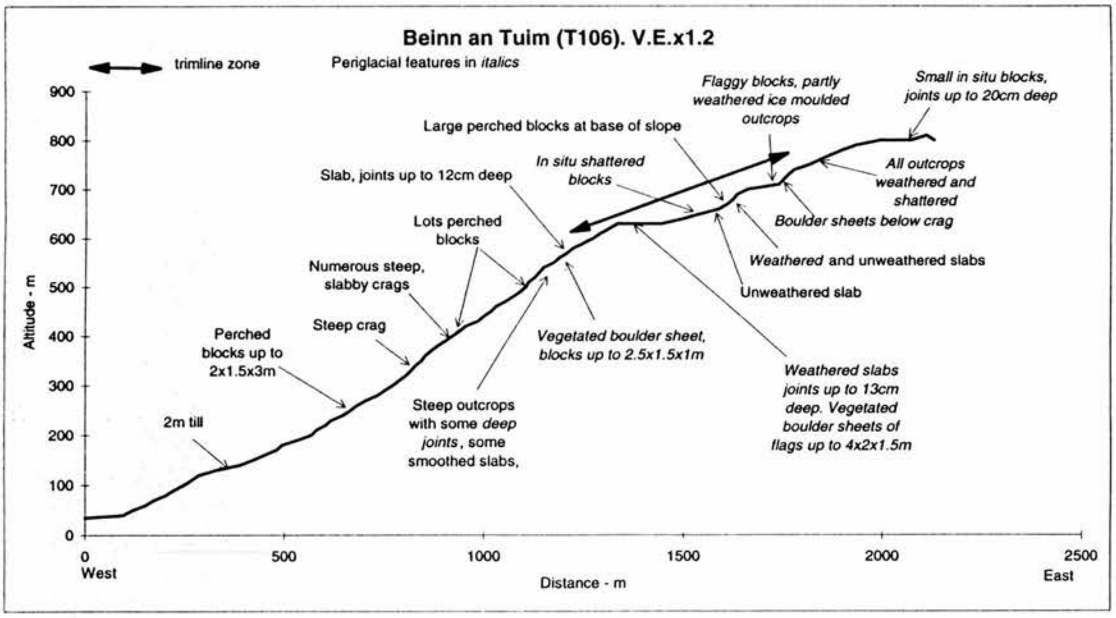


Fig 3.6 Transects showing trimline evidence on two slopes with broad trimline zones.

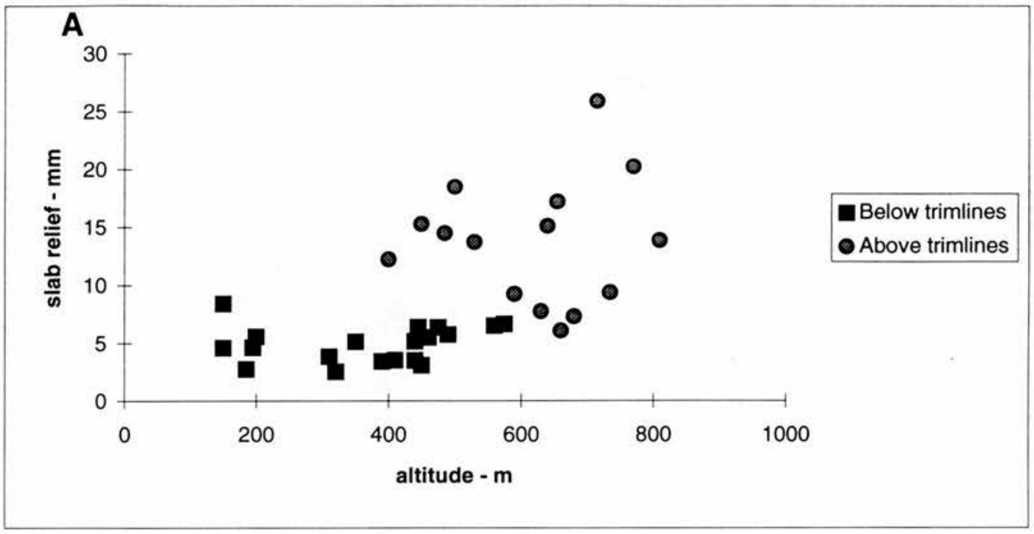
of Glen Dubh Lighe and the NW side of Loch Shiel. Trimlines altitudes from these areas are ~100m, ~10m and ~50m higher, respectively, than those immediately adjacent to them.

The evidence on low summits, ridges and cols provides additional evidence to support the trimline altitudes found on adjacent slopes (Tables A.4.2 and A.4.3 in Appendix 4). In many instances summits and cols below adjacent trimline altitudes show fresh glacial evidence, and those above show clear periglacial features. However, the evidence in these sites is not always clear. Summits below the altitude of adjacent trimlines such as Meall a'Choire Chruinn (S12) (illustrated in Fig 3.8) and also Sgorr nan Cearc (S9), commonly have frost shattered outcrops and frost shattered debris, whereas the valley side slopes immediately beneath them have extensive smoothed, unweathered slabs. Several cols at various altitudes have a mixture of glacial and periglacial evidence, which cannot easily be interpreted. It is possible that these difficulties are a result of the topographic changes associated with summits and cols. Hence the evidence from these locations is given less emphasis than trimline altitudes from slopes and spurs.

3.5.2 Quantitative approaches

Slab Relief

34 slab relief measurements were taken from typical slabs below, around and above the trimline altitudes on 11 different slopes. The results are listed in Appendix 4, Table A.4.4. This table shows that the mean and standard deviation of slab relief at points on a traverse up a slope is always greater above trimlines than below, with the actual values being influenced by local geology. In general, slabs below trimlines have mean and standard deviation reliefs of less than 6mm, whereas slabs above trimlines have values of more than 12mm (Fig 3.7a). Despite the influence of geology there is very little overlap in the values obtained from above and below trimlines on different rock types, although the higher values obtained below trimlines are all from fissile micaceous schist slabs, and the lower values obtained above trimlines are usually from slabs of massive gneisses. Slab relief shows a closer relationship with position in relation to the trimline than it does with altitude (Fig 3.7b). This method thus provides a useful quantitative approach to measuring trimline altitudes in Western Lochaber. Few slabs below 300-400m are weathered, so the method is most useful in the mid and upper reaches of the former glaciers.



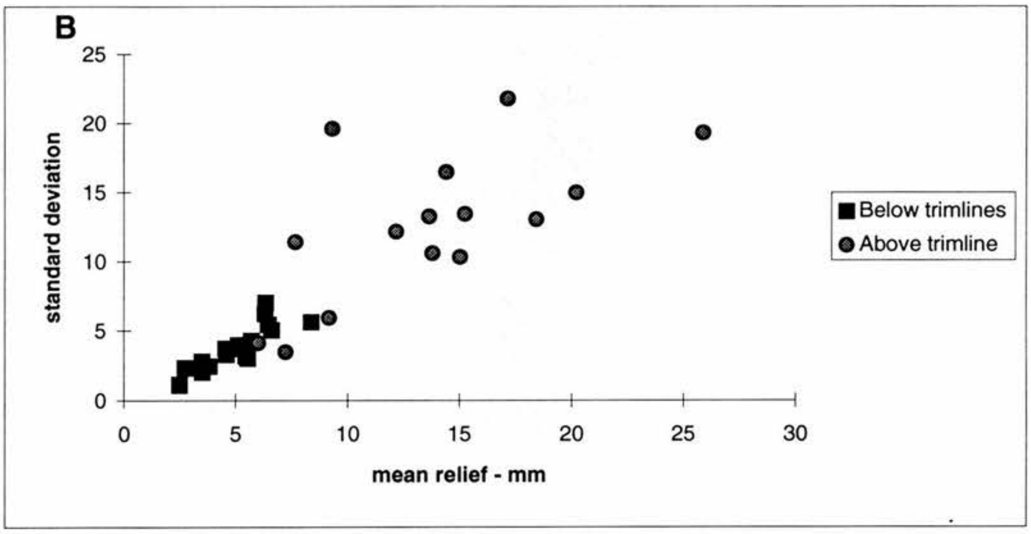


Fig 3.7 Slab relief above and below trimlines - mean relief of measurements at 2cm intervals over 70cm on planar slab outcrops.

A. Variation of mean slab relief with altitude.

B. Variation of mean slab relief with standard deviation.

A. B.

Fig 3.8 Plates showing example of contrasts in bedrock weathering on steep glen walls and gentle summit slopes directly above, 630m on Meall a'Choire Chruinn (S12).
Tape measure indicates scale.
A. Smoothed unweathered slab on steep glen sides.
B. Frost shattered and weathered bedrock outcrops on gentle summit slopes.

Depths of open joints

Table A.4.5 in Appendix 4 shows 189 measurements of joint depths in outcrops on 64 different slopes; on 51 of these slopes measurements were taken from both below and above the trimline. Joint depths are of variable use in providing additional trimline evidence; of the latter 51 slopes, 17 show an increase in depths above the trimline, although this increase was often small, and 34 show no trend with respect to the trimline altitude. Fig 3.9a shows that the overall distributions of mean joint depth with altitude in Western Lochaber are similar above and below trimlines. This suggests that mean joint depth is not a good indicator of position with respect to the trimline. As this could be due to geological variations within the area, Figs 3.9b-g show the frequency distribution of joint depths at different altitudes on some of the slopes. Fig 3.9b shows that, in some cases, deep joints are only present above trimlines. Elsewhere, the relationship is less clear (Fig 3.9c). In general, shallow joints of 0-5cm are common at all altitudes, regardless of position with respect to trimlines. Deep joints are usually confined to gentle summit and ridge top slopes above trimlines; they are not present on steep glen sides above trimlines (Figs 3.8, 3.9e). Field observations suggested that joint depths were influenced by slope angle, rock outcrop surface angle and geology. Where possible,

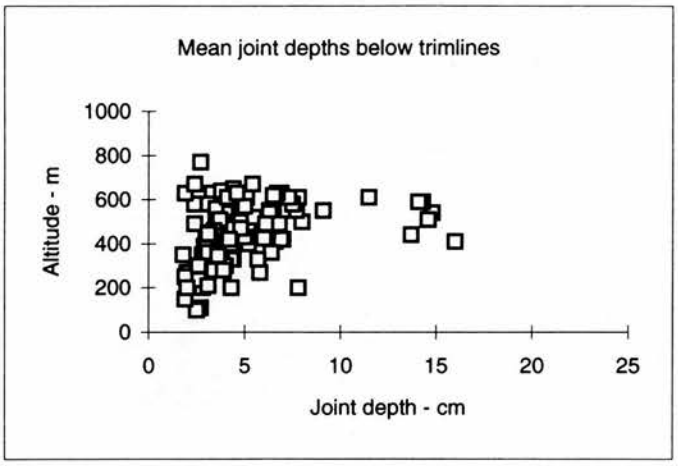
depths were recorded from outcrops selected to keep these factors constant on each slope. Overall, joint depths show a strong relationship with these factors.
The relationships are:
- nature of the rock outcrop. Fig 3.9d shows the distributions of joint depths taken from two different outcrops below the trimline on Sgurr a' Bhuic at 440m (S7, T38). The joints in the slab are mostly shallow, whereas there are numerous deep joints in the steep face outcrop. An identical pattern exists above trimlines; there are abundant relict smoothed slabs above trimlines in which there are no deep joints. It is estimated that the joint depth measurements on 24 of the 51 slopes were affected by the predominant form of outcrop on the slope (i.e. is slabs or free faces). Joint depths on slabs are shallow at all altitudes, and deep on free faces.

- slope angle. Fig 3.9e shows the distribution of joint depths from outcrops immediately above and below the break in slope between the steep glen sides and gentle summit slopes at 630m on Meall a' Choire Chruinn (S12, T28). This is below the trimline altitude. The outcrop on the steep glen side has shallow joints whereas that on the summit slopes shows evidence of frost shattering and has deeper joints (see Fig 3.8). On slopes of less than 15^{o} outcrops always have deeper joints than those on steeper glen sides immediately below. It is estimated the measurements on 16 of the 51 slopes were affected by these changes in slope angle. This pattern may be related to free water drainage on steep slopes reducing water availability for freeze-thaw action.

- geology. Lithology and structure have a large influence on joint depths , so it is unwise to compare depths from different rock types. Fig 3.9f illustrates an influence of local geology; it shows the distribution of joint depths taken from outcrops of psammitic country rock and a fine grained acid intrusion at the same altitude on Meall a' Bhragaid (T50).

- random factors. At any one altitude there can be a wide variety in the degree of weathering of different outcrops. This is illustrated in Fig 3.9g which shows joint depth frequency histograms from adjacent outcrops showing different degrees of weathering on 2 different slopes.

The main problems with the use of joint depths in Western Lochaber are firstly the widespread distribution of relict Devensian smoothed slabs above the LLS trimlines. Since these slabs have smooth surfaces resistant to frost penetration, there are few outcrops above trimlines showing deep joints. Secondly, the presence of deep joints above trimlines is largely controlled by slope angle; deep joints are normally restricted to gentle summit slopes and are not widespread on steep glen sidewalls.



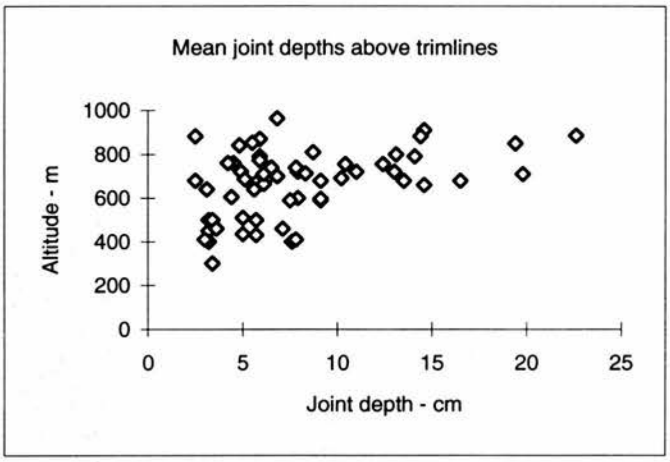
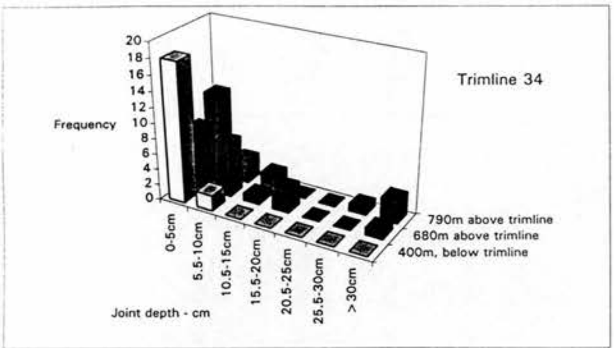
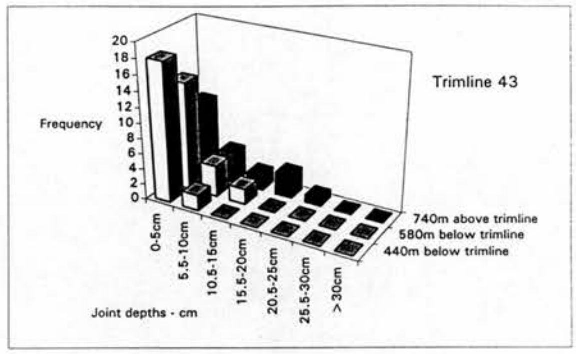


Fig 3.9a Variation of mean joint depth with altitude for bedrock outcrops below and above trimlines.





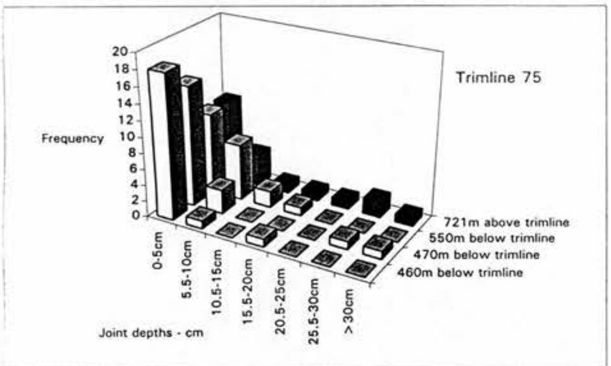


Fig 3.9b Frequency histograms of joint depths on outcrops below and above trimlines, examples showing deeper joints above trimlines.

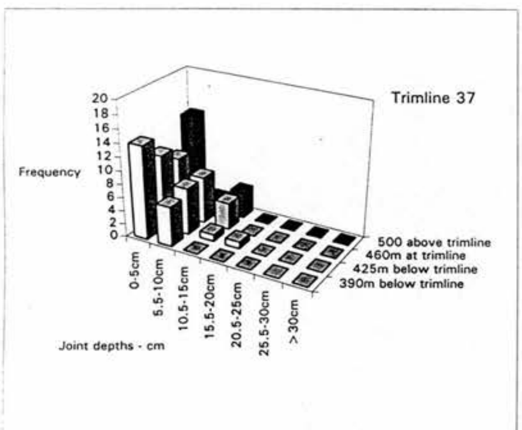
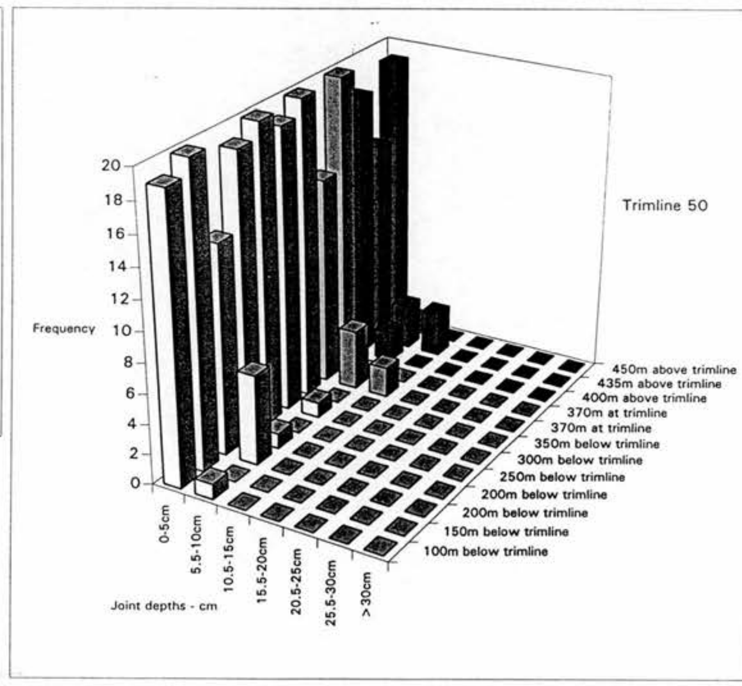
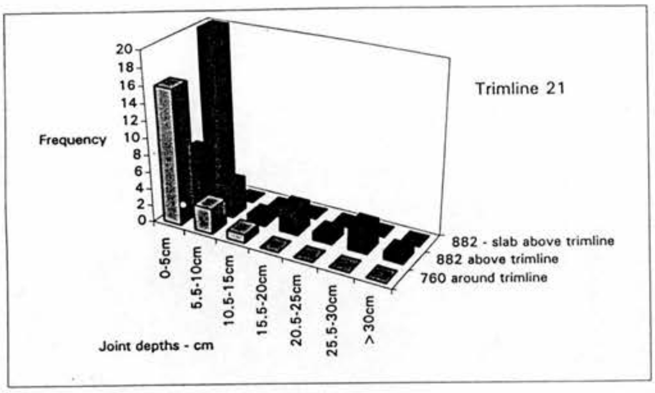


Fig 3.9c Frequency histograms of joint depths on outcrops below and above trimlines. Examples showing no increase in joint depths above trimlines





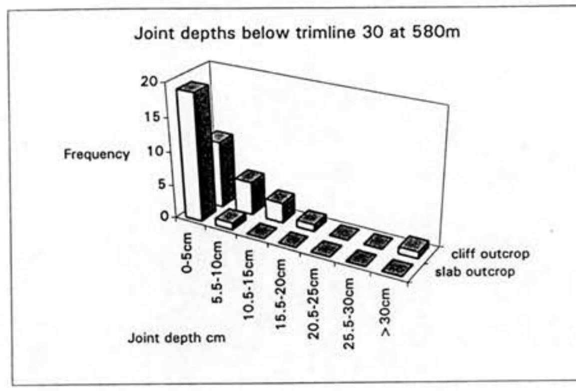
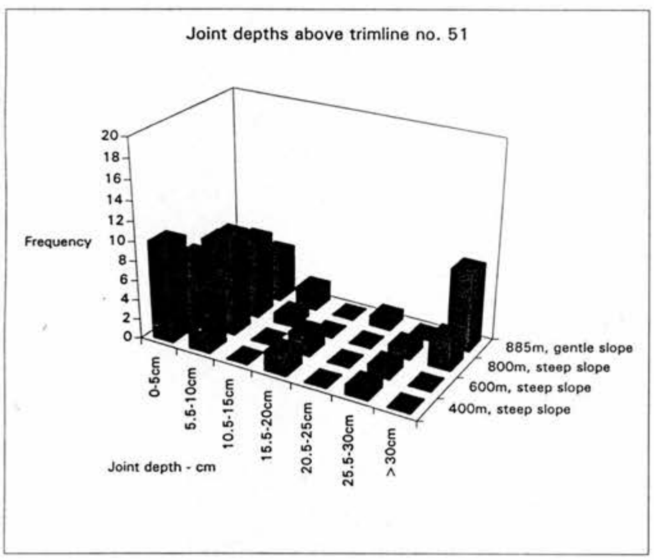


Fig 3.9d Frequency histograms of joint depths on different types of bedrock outcrop, examples showing shallow joints on slabs, and deep joints on cliffs.



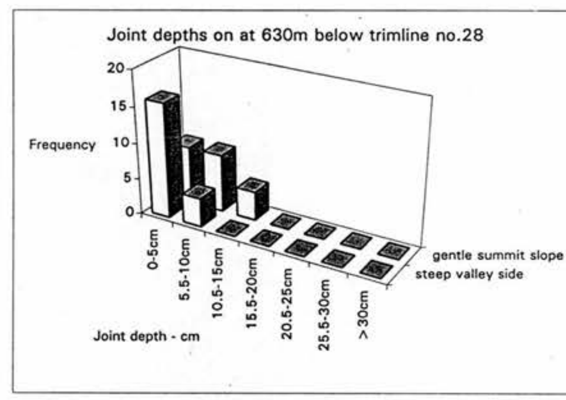


Fig 3.9e Frequency histograms of joint depths on different outcrops, examples showing deep joints on outcrops on gently sloping ridge tops, and shallower joints on outcrops on steeper valley sides below.

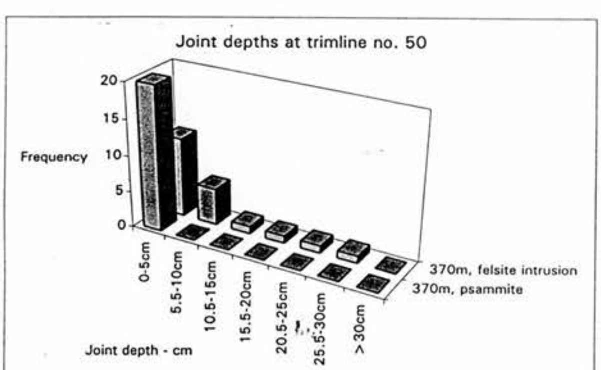


Fig 3.9f Frequency histograms of joint depths on outcrops of different lithology at the same altitude, on the same slope.

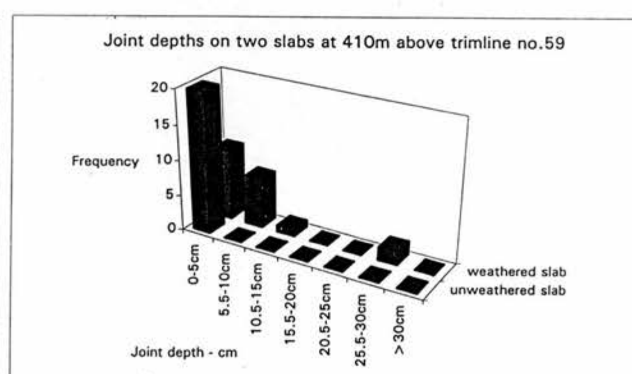


Fig 3.9g Frequency histograms of joint depths on adjacent slab outcrops, showing a high degree of local variation in surface weathering.

3.6 The reliability of trimline evidence

It is argued here that the trimlines represent the upper surface of ice during the LLS glaciation for several reasons. Firstly, the trimlines are locations of real changes in the nature of the landforms present on slopes. In the clearest trimlines, such as those shown in Fig 3.5, sudden changes in the distribution of glacial and periglacial features occur over altitudinal distances of as little as 10m.

Secondly, the trimline altitudes are easily explicable by the distribution of a former ice cap as, with only three exceptions from 110 values, they form a sensible regional pattern showing ice flowing consistently away from the mountains and down troughs. Cross- and down- trough ice surface profiles are all plausible (Fig 3.4), and consistent from glen to glen. The trimline altitudes are highest in the north east of the area, the central mountain belt in Ardgour, and the higher mountains in the north and north west. The altitudes decline towards the loch troughs in the west and south.

Thirdly, the clarity of the trimlines mapped conforms well to the pattern predicted earlier in this chapter on the basis of theoretical considerations:

- glacial regime. Many of the slopes on which trimlines were very difficult to identify are located in the vicinity of former ice sheds. 27% of the mapped slopes close to the likely former ice sheds do not have clear glacial evidence up to the trimline altitude. Trimlines are easier to locate further downglacier where glacial erosion has been efficient. Additionally, the slopes in SW Glenfinnan, W Glen Dubh Lighe and NW of Loch Shiel where the evidence suggested anomalous trimline altitudes may be attributed to the location of the massifs near the ice shed of the reconstructed ice cap. This inaccuracy towards the former ice sheds is likely to be the main source of error in the reconstruction, and the form of the ice surface in these areas is less certain.

- slope angle. Trimlines are difficult to locate on very steep slopes of 40^{o} or more such as on Beinn a' Chaorain (T26) as mass movement processes are active. This means trimlines are difficult to pinpoint accurately; 4 of the trimlines located on the 12 steepest slopes can only be identified to an accuracy of >100m. On gentle ridge top and plateau slopes of less than 15^{o}, such as on Sgurr na Greine (T30), weathered and shattered outcrops are found below the trimline. Slope angles intermediate between these two values tended to show clear trimline evidence.

- geology. Where the planes of weakness in the underlying bedrock are narrowly spaced, frost shattered debris is found at all altitudes. This is the case on many slopes in the north west of the study area, which are underlain by fissile, micaceous semipelitic and pelitic schists. 67% of the slopes on these micaceous schists have periglacial features below the trimline altitude. The trimlines in lower Glen Moidart are located on this bedrock, as well as

being at a low altitude, which may explain the lack of clear evidence here. Where the planes of weakness are widely spaced, such as on Sron Liath (T99), smoothed slabs are well preserved above the former glacial limit and there is little frost shattering, and this produces blocky rather than flaggy debris. In both cases trimline altitudes are hard to locate.
- altitude. Periglacial features are not well developed below 300 - 400m in Western Lochaber. For this reason, just 1 of the 14 trimlines below 400m shows firm trimline evidence with clear periglacial evidence above the trimline. Periglacial features are best developed above 600m; 98 % of the trimlines at or above 600m have clear periglacial evidence.

3.7 Conclusions

1. The consistency, number and pattern of trimlines mapped suggests that they represent the surface of a former ice cap of Loch Lomond Stadial age.
2. No single feature can be used to identify trimlines, but rather assemblages of features at different altitudes must be used due to the wide range of natural variations.
3. Measurement of the degree of weathering along the foliation planes on a typical smoothed slab is an effective means of locating trimlines where relict smoothed slabs from previous glaciations have been preserved above the line.
4. Measurements of the depths of open joints in rock outcrops do not always give a reliable indicator of the location of trimlines in an area where relict smoothed slabs are widely preserved above trimlines, and where ice often extended above steep glen sides onto gentle summit slopes above.
5. The clearest trimlines are obtained

    - away from the former ice sheds.
    -on moderately steep slopes of 25^{o} -35^{o} with a constant slope angle at right angles to the direction of ice flow.
    - on slopes with exposed bedrock.
    - on slopes where the bedrock is moderately resistant to frost action, such as psammitic schists and gneisses.
    -above 400m.

Chapter 4 - Seismic evidence

4.1 Aim

In this chapter the Quaternary sedimentary record in Loch Linnhe and the Firth of Lorne is investigated, by means of a seismic survey, in order to provide additional evidence with which to reconstruct the changing environmental conditions in Western Lochaber through the late Quaternary. The inner sea lochs are sediment traps and contain a record of changing terrestrial environmental conditions in a region where the adjacent land area has undergone strong erosion and generated a large volume of sediment. With the exception of studies by Boulton et al. (1981), Deegan et al. (1973), and Dix (cited in Benn 1991), there have been few investigations of the deposits in the inner sea lochs, which offer the potential to link the terrestrial and continental shelf records.

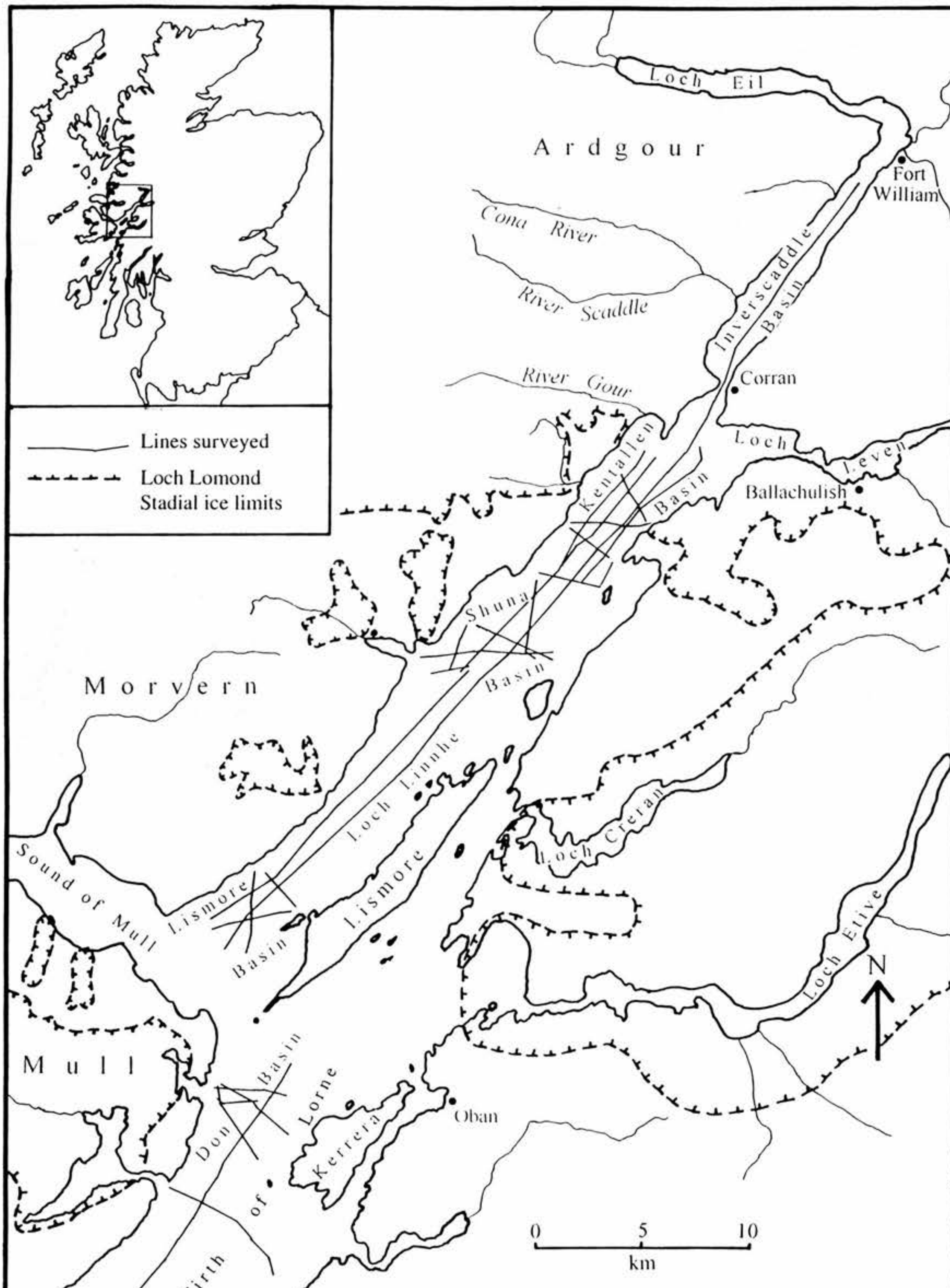


Fig 4.1 Lines surveyed in Loch Linnhe and the Firth of Lorne.

There is terrestrial evidence that Loch Lomond Stadial ice cap outlet glaciers terminated within the Loch Linnhe and the Firth of Lorne areas (Fig 4.1) (Thorp 1986, Gray and Brookes 1972). If so, these limits should be associated with evidence on the floors of the sea lochs and the sea floor evidence may augment available terrestrial evidence. As the areas surveyed lie both outside and inside the presumed former LLS ice limits (Fig 4.1) (Thorp 1986), it was anticipated that the results would both test existing hypotheses concerning the location of LLS glacial limits based on terrestrial evidence, and provide additional evidence concerning the distribution of LLS glacigenic sediment.

The lochs are mostly between 60 and 100m in depth, suggesting that the sedimentary record should not have been strongly affected by tidal scour. Furthermore, as one of the widest Scottish sea lochs, Loch Linnhe offers the best possibilities for obtaining transverse survey lines and a fine grid coverage of the survey area.

4.2 Methods

4.2.1 Seismic stratigraphy

Seismic methods generate an acoustic pulse from the survey vessel which passes through the water and ground beneath the vessel until it is reflected. Transmission of the wave is density dependent, and is most efficient through dense media. Reflection occurs when the wave reaches interfaces between media of different acoustic properties. The frequency distribution of energy in the reflected wave differs from that in the incident wave, and the difference is dependent on the acoustic properties (the impedance, which is related to density) of the rock. If there is a large difference in the density of two adjacent strata a strong reflector appears on the seismic record. Reflectors become weaker with depth due to the absorption of energy.

The interpretation of seismic records rests on the principle that the different acoustic impedances of different lithologies have a different effect in changing the frequency spectrum of the original wave.

If the input wave is constant, then similar lithological sequences will generate similar reflected waves.

Different seismic stratigraphic units are separated on the basis of their geometric configuration and the continuity and the amplitude of the reflectors (Fig 4.2). These are correlated between intersecting survey lines to build a stratigraphy based on acoustic properties. The stratigraphic position, physiographic location, geometric organisation, strength and continuity of these reflectors, and a consideration of sedimentological theory may also suggest the types of sediment represented by the seismic units, and the environmental conditions at the time of deposition. Using the previously established regional chronology of environmental change during the late Quaternary (see Section 1.4), ages for the different units can be suggested.

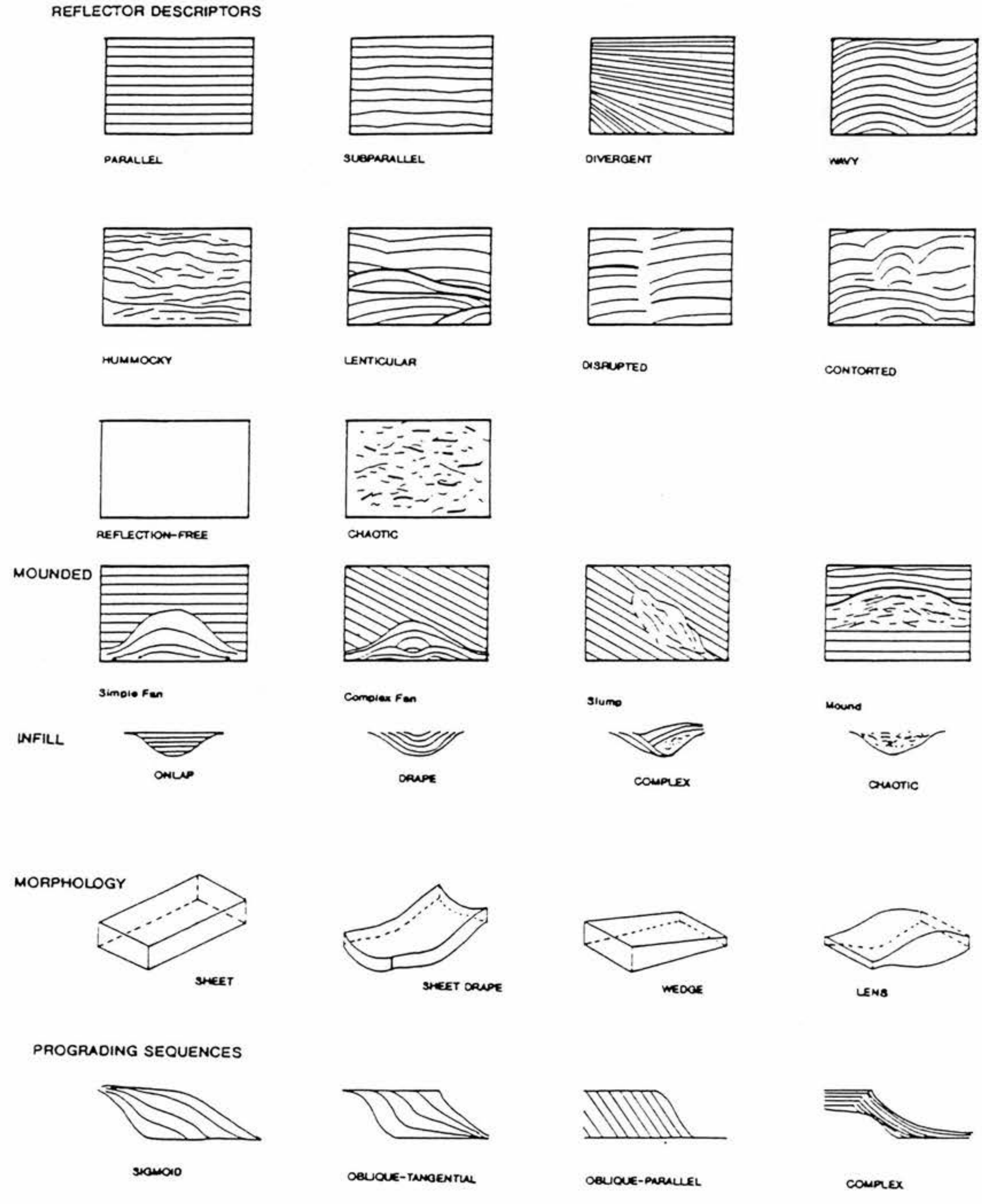


Fig 4.2 Reflector discriptors and seismic unit facies forms, from Stewart (1991), (after Mitchum et al. 1977, and Sangree and Widmier 1977).

Limitations in seismic stratigraphy result from the following problems (based on Stewart 1991):-

1. Resolution. Seismic records can only show the gross sedimentary structures, and important variations in physical properties over small distances may not be detected. In this study the resolution is likely to be 3-10m, as maximum resolution is twice the wavelength of the reflected wave.

2. Acoustic impedance. Layers with a high impedance may obscure the underlying record and adjacent units may not be distinguished if their acoustic impedances are similar. Also, gradual vertical changes in sediment density due, for example, to compaction, may result in the crossing of a threshold in acoustic impedance and produce an apparent reflector on the seismic record which may not correspond to a lithological change.

3. Resonance. The seismic signal can be reflected several times between any two reflectors before it is detected, which results in spurious 'multiple' reflectors below the 'real' ones.

4. Distortion. Records may be distorted, for example by refraction from inclined surfaces or point sources, if gas in the sediments transmits energy inefficiently, blanking out the record from below, or if the velocity of sound is significantly different between different media.

5. Reproducibility. The seismic stratigraphy obtained results from a subjective interpretation by the researcher about which reflectors are important and which are spurious.

6. Location. The limitations of the positioning systems used to fix lines may be insufficient for detailed correlation between survey lines, especially where units have rapidly changing spatial characteristics.

7. Generic assumptions. It is assumed in this method that seismic facies do not change their gross sedimentological properties over the area studied, that similar seismic signatures represent similar sedimentary sequences, and that different sediment sequences have different seismic signatures.

These limitations mean that seismic reflection profiles are interpreted as indicators of the large scale organisation of beds, rather than an accurate representation of detail. It is widely recommended (e.g. Carlson 1989, Stewart and Stoker 1990) that sedimentological controls on seismic interpretations are provided by cores located on seismic traverses. Deep drilling equipment was not available in this study, hence the seismic interpretations are not constrained by lithological evidence and must remain provisional.

4.2.2 Survey details

On 24th - 26th August 1992 a seismic survey was made in Loch Linnhe and the Firth of Lorne. The R.V. Calanus, a 20m research vessel, ran at a surveying speed of 4 knots towing a multi-tip sparker and Teledyn CH1-4 hydrophone. The energy level used was 500J with a firing interval of 0.8 seconds. Recordings were made in the 320-1000Hz range. Positions were fixed by G.P.S. navigational equipment, which has an accuracy of +/- 50m, and recorded every 5 minutes. The survey commenced with a series of longitudinal lines in Loch Linnhe. The location of the later lines were chosen so as to link the sedimentary basins already found (Fig 4.1). 171 km were profiled. In the following discussion, indices of depth have been converted from two-way travel time in milliseconds to depths in metres, and this has been based on a velocity of sound in water of $1490 ms^{-1}$ and in soft sediment of $1800 ms^{-1}$ (McQuillin and Ardus 1977). This conversion is likely to provide a minimum estimate of sediment thicknesses.

4.3 Results

Most of the seismic records obtained are of a high quality, with different sedimentary units and the sediment/ bedrock interface clearly visible. In outer Loch Linnhe, for example, there is a large basin 280m deep containing sediments which are locally more than 160 m thick. Some of the records suffer from acoustic blanking, mainly in the part of Loch Linnhe between Lismore and Kingairloch. This is thought to be due to energy absorption by biogenic gas in the sediments.

In the following sections the different seismic facies present in Loch Linnhe are identified, likely environments of deposition are suggested for each, a seismic stratigraphy for each basin is constructed, and the implications of each stratigraphy discussed.

4.3.1 Seismic facies in Loch Linnhe

The following seismic facies are distinguished in the Loch Linnhe on the basis of the configuration, continuity and amplitude of reflectors. Examples of each facies appear in Fig 4.3.

Facies A - draped reflection pattern with internal reflectors subparallel to the base of the facies

Facies B - infill reflection pattern with subparallel, subhorizontal internal reflectors. There are intermediate patterns between facies A and B.

Facies C - a sediment mass banked up against bedrock. This is either wedge shaped (Fig 4.3c) or hummocky (see Fig 4.5c, fix 31), often with a strong, inclined surface reflector and containing strong, inclined subparallel internal reflectors.

Facies D - this often has an irregular surface profile, with incoherent or hyperbolic point source internal reflectors.

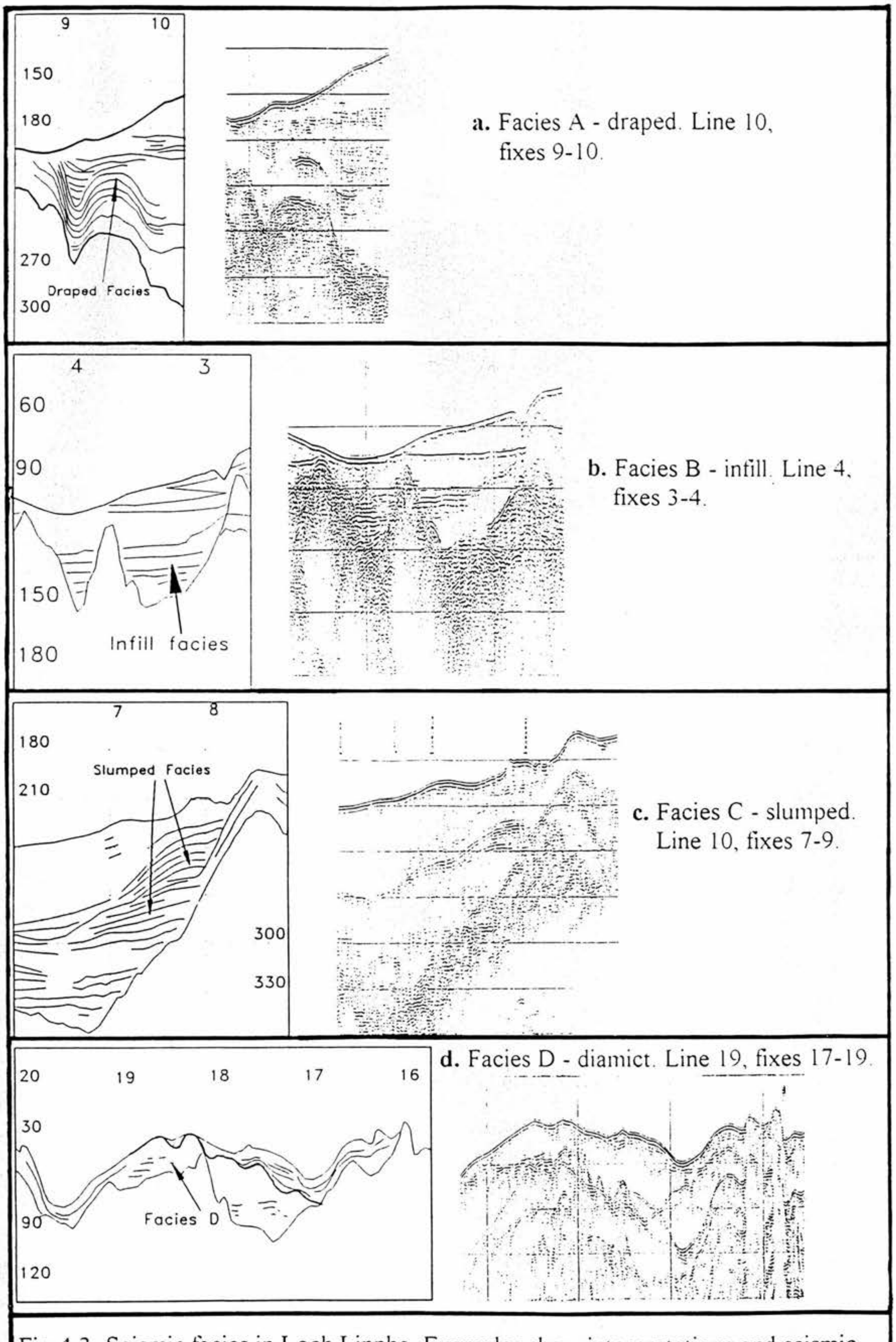


Fig 4.3. Seismic facies in Loch Linnhe. Examples show interpretations and seismic originals. Vertical units are Two-way travel time (ms), horizontal units are fix numbers.

Facies E - A unit with a strong and highly irregular surface reflector, and dense incoherent reflectors close to the surface. Facies E is present at the base of the seismic records shown in Figs 4.3a-d, and Figs 4.4 - 4.7. 'Ghost 'reflectors are often found above and below the main surface reflector (see Fig 4.7, between fixes 4 and 5).

4.3.2 Interpretation of the seismic facies

The following interpretations are partly based on those of Mitchum et al. (1977) and Syvitski (1989).

Facies A is interpreted as a deposit from a surface water plume with high suspended sediment concentrations. The arrangement of the internal reflectors suggest deposition in a stable setting. Such plumes are typically produced by turbid freshwater inputs into the marine environment.

Facies B shows an infill internal structure. These structures are produced in a higher energy environment than Facies A, for example by traction currents moving sediment over the loch bed or by the slumping of unconsolidated sediments from topographic highs.

The geometry of Facies C suggests that it is a slumped unit. In places it shows several strong subparallel internal reflectors which may indicate the repetitive nature of slumping and debris flow.

Facies D is interpreted as a diamicton unit. The irregular surface profile and lack of internal structure is typical of glacigenic deposits and the internal hyperbolic reflectors probably result from boulders within it. Proximal glacimarine and glacial till units cannot normally be distinguished on the basis of seismic evidence (Stewart and Stoker 1990, Davies et al. 1984); hence this facies is interpreted as a glacigenic diamicton, deposited close to, or in contact with, glacier ice.

Facies E is interpreted as bedrock. The strong, continuous surface reflector indicates a large density contrast with the overlying sediments. Occasionally patterns are visible below this reflector which may represent structures within it. The 'Ghost' reflection patterns can be explained by a consideration of the geometry of the incident and reflected energy signals. The seismic signal spreads out radially to form a cone. If the bed has an irregular surface profile, peaks not directly below the track line will also reflect back the energy pulse. As the signal has travelled a different distance to and from these adjacent peaks, it will be received by the hydrophone at a different time to reflectors in the same plane directly below the ship. Thus the ghost reflections which appear on the seismic record before or after similar peaks on the main bedrock reflector may to be from ridges which cross the track line. Ridge structures can be discerned where one of the limbs of the peak in the ghost is in line with the limb of a peak in the main reflector.

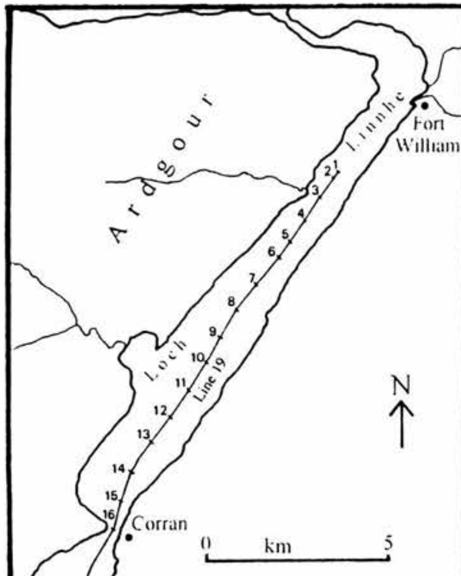
4.4 The seismic stratigraphy in Loch Linnhe

The bedrock reflector shows an irregular, high amplitude relief. The bedrock surface displays numerous troughs and basins which have influenced the distribution of the sediment masses in the loch (e.g. Fig 4.5c). Bedrock basins show an irregular basal topography and attain considerable depths, clearly showing the results of intense glacial erosion. There are five main basins in the area surveyed. The Inverscaddle basin lies in upper Loch Linnhe and attains a maximum surveyed depth of >375ms (>~308m). South of the Corran narrows are the smaller and shallower Kentallen and Shuna basins. Line 10 shows that these basins separated by a small bedrock ridge on the sea bed. The Lismore basin is a deep basin in outer Loch Linnhe, attaining a depth of 351ms (~280m). The survey lines show several small basins in the Firth of Lorne. The largest of these is the steep-sided Don basin which is 418ms (~325m) deep. A seismic stratigraphy for each basin is now constructed, distinguishing different units on the basis of the continuity, amplitude, and geometric configuration of the reflectors, and correlating these between seismic profile lines and, where possible, between basins.

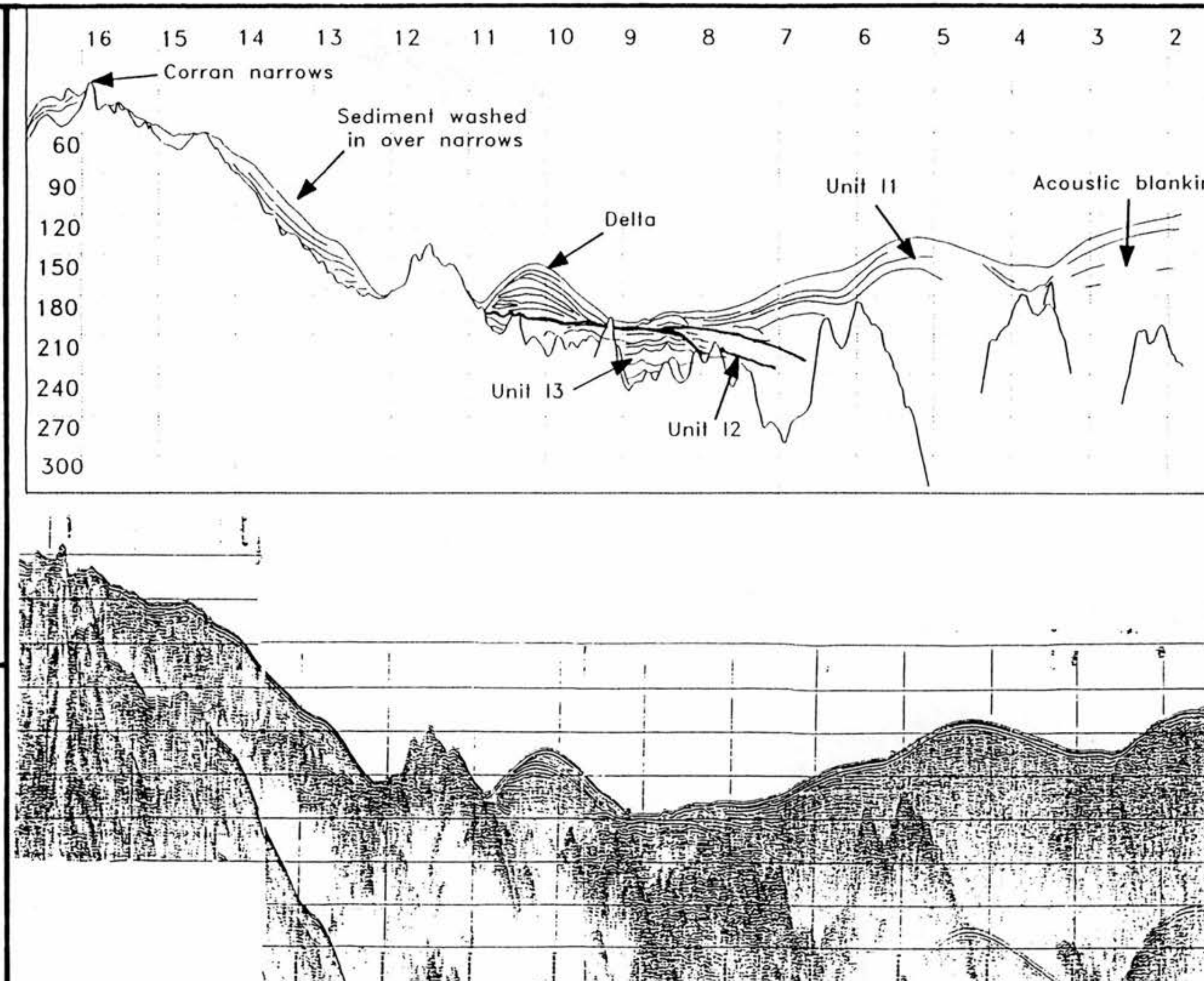
4.4.1 Inverscaddle basin

The Inverscaddle basin lies in Loch Linnhe between Corran to within 5km of Fort William. There is probably more than 165m of sediment in the deepest part of this basin, although detail at this point is obscured by acoustic blanking. Figure 4.4 and Table 4.1 summarise the seismic stratigraphic units in the Inverscaddle basin, based on the seismic reflection characteristics of line 19.

Unit I1 is the uppermost and thickest unit, and is widely distributed on the loch floor. It shows the characteristics of facies A, sometimes grading into facies B at topographic highs. This suggests Unit I1 was deposited from a surface water plume with high suspended sediment concentrations, with some resedimentation by bottom currents. The thickness and lateral distribution of this unit suggests three main sediment sources. Firstly, sediment has been washed in by tidal currents and/ or periodic deep water flows over the Corran narrows. Secondly, sediment has been discharged from the river Scaddle, the mouth of which is marked by a large mound with a deltaic structure which has a volume of approximately $1.4 \times 10^7 m^3$.



a. Location of survey lines, and fixes shown in Fig 4.4b.



b. Line 19 - seismic stratigraphic units and seismic original.

Fig 4.4 Seismic stratigraphy of the Inverscaddle basin.
In Figs 4.4 - 4.7 all vertical units are two-way travel time (ms), and horizontal units are fix numbers.
In the interpretations, thick lines represent reflectors separating different units.

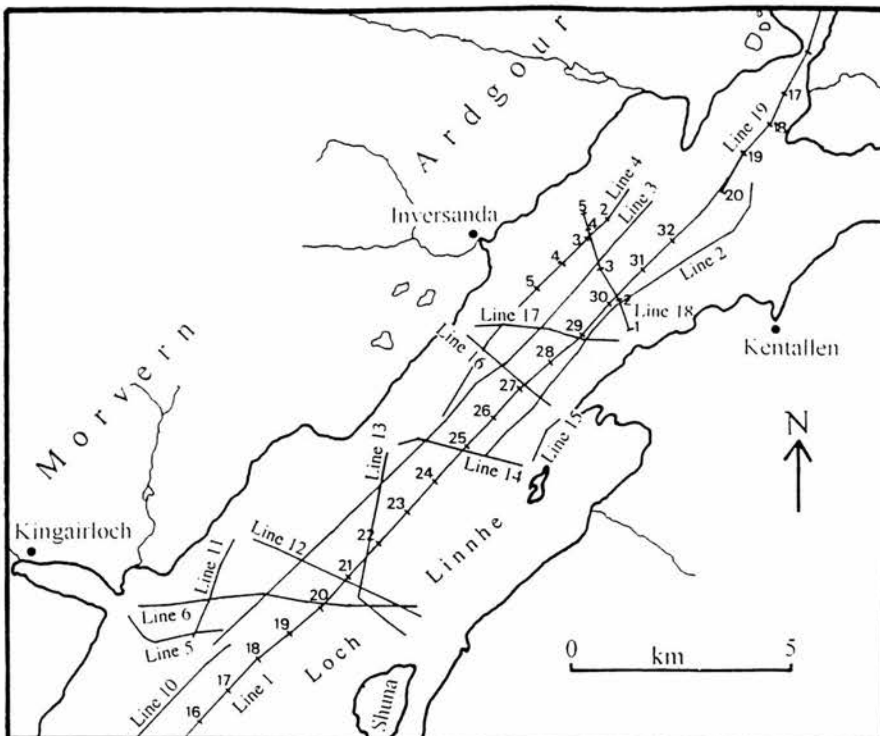
Table 4.1. Units in the Inverscaddle basin.

Unit	Surface reflector	Internal reflectors	Geometry	Thickness	Environment of deposition	Possible age
I1	strong	strong, sub parallel	draped, slumped, smooth mounds	>60ms (>54m)	bottom currents, deltaic progradation, debris flows	Holocene
I2	medium	incoherent, hyperbolic	infill?	up to 20ms (18m)	in situ, ice contact?	L.L.S. advance?
I3	strong	strong, sub parallel	draped in basins	up to 49ms (44m)	suspension fallout, glacimarine?	lateglacial deglaciation? (/ L.L.S)?

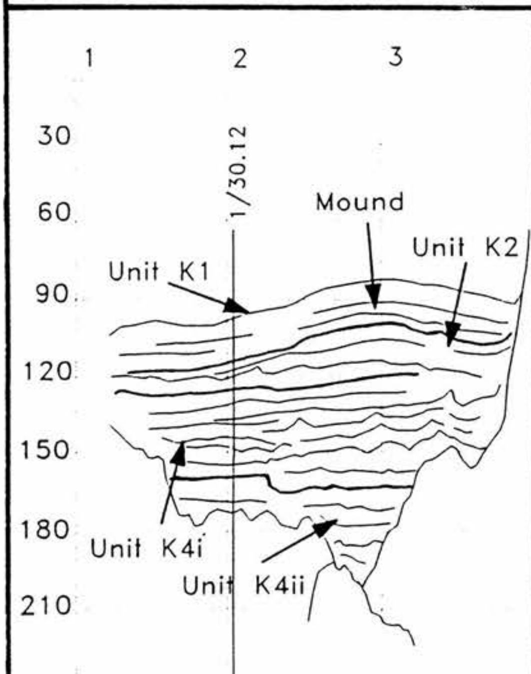
Thirdly, the unit thickens to the north of the Loch where three more large rivers drain into Loch Linnhe, and the smooth surface reflector suggests the sediment unit may be a distal delta. The manner in which these three locations correspond with contemporary sediment sources, the stratigraphic position of the unit and the inferred processes of deposition all suggest that this unit represents sedimentation under marine conditions similar to those of today. The thicknesses of distal deltaic sediments suggest that the unit has accumulated throughout the Holocene, as it is unlikely these volumes of sediment could have accumulated in the short time between deglaciation and the early Holocene. The smooth surface reflector suggests that the unit is being deposited at present. For these reasons Unit I1 is interpreted as a Holocene unit.

Unit I2 is present in one location, and the characteristics of the unit are not shown clearly on the seismic record due to acoustic blanking in the northern part of the basin. It may be composed of facies D, and if so may represent a glacigenic diamicton deposited during the LLS glacial advance.

Unit I3 shows characteristics of facies A and B, and is interpreted as a unit deposited both by suspension fallout and debris flows in the relatively high energy glacimarine environment. This may have been deposited during deglaciation following the Lateglacial Maximum (LGM).



a. Location of survey lines, and fixes shown in Figs 4.4b and 4.4c (overleaf).



b. Line 18 (transverse line) - seismic stratigraphic units and seismic original.

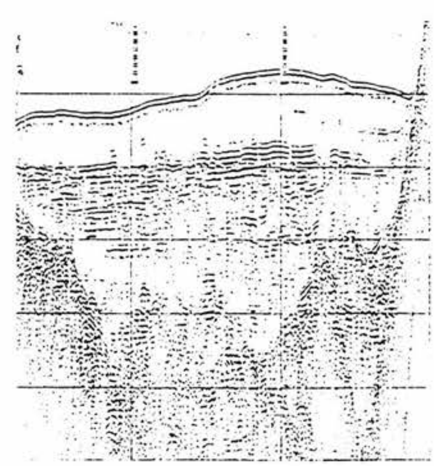


Fig 4.5 Seismic stratigraphy of the Kentallen and Shuna basins.

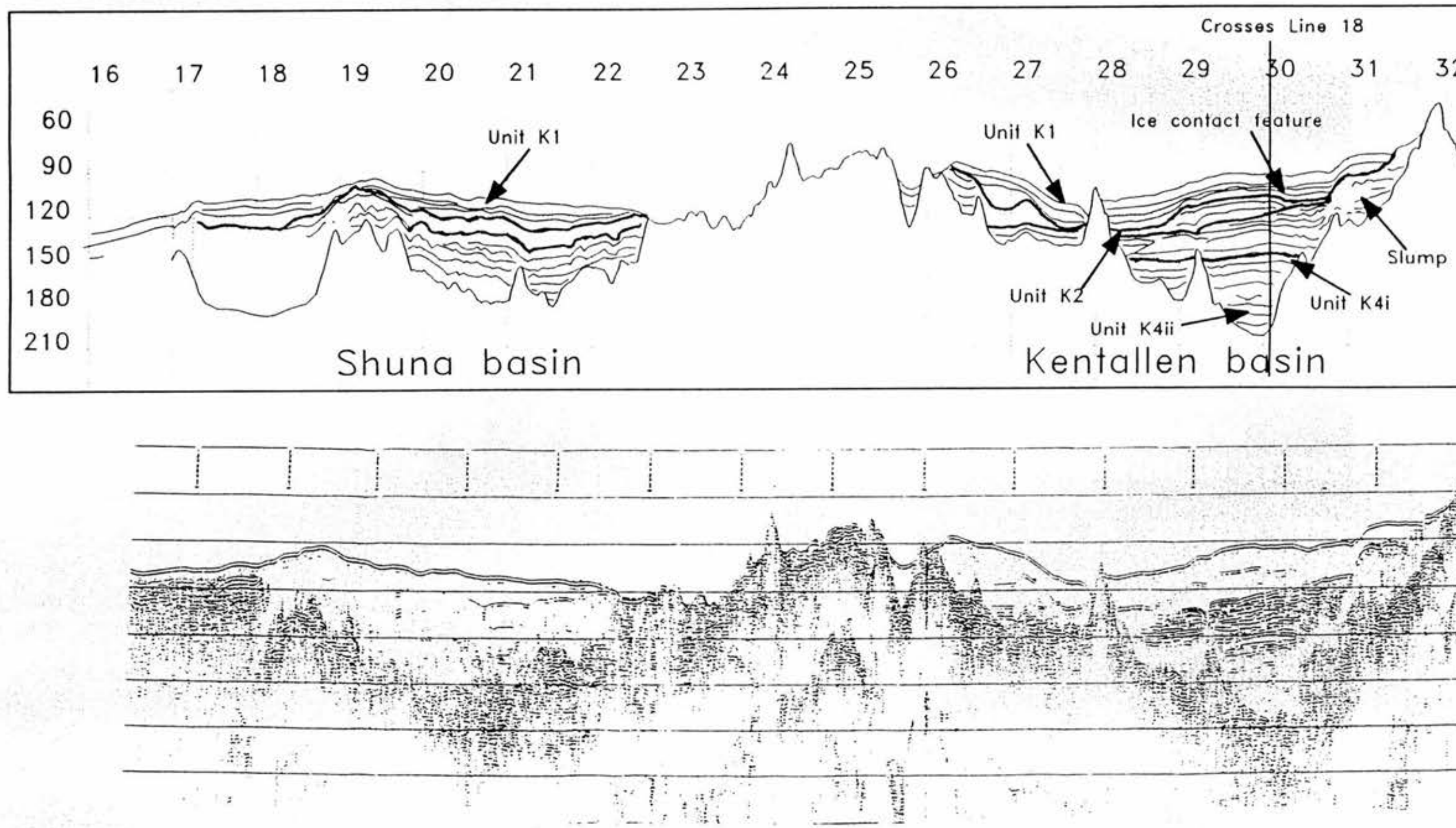


Fig 4.5c Line 1 (longitudinal line) - seismic stratigraphic units and seismic original.

4.4.2 Kentallen and Shuna basins

These basins lie south of the Corran narrows. The Kentallen bedrock basin is 215m deep, and the maximum thickness of sediment is more than 147m. The deepest bedrock basin in the Shuna basin is 191m and this contains sediments up to a maximum thickness of 108m. There is a grid coverage of clear seismic records in these basins, and as a result the seismic interpretations here are the most secure of the basins surveyed. Figure 4.5 shows the typical arrangement of facies in the Kentallen and Shuna basins, and seismic stratigraphic correlations between them. The characteristics and paleaoenvironmental interpretations of the units are summarised in Table 4.2.

Table 4.2 Units in the Kentallen and Shuna basins.

Unit	Surface reflector	Internal reflectors	Geometry	Thickness	Mode of deposition	Environment of deposition	Possible age
K1	strong	weak, sub parallel	infill, some draping in basins	up to 38ms (34m)	suspension fallout, bottom currents	non-glacial	Holocene
K2	strong	incoherent, hyperbolic	irregular constructional mounds	up to 30ms (27m)	in situ	ice contact	Loch Lomond Stadial
K3	weak	weak, sub parallel	basin fill and mounds in Kentallen basin, draped in Shona basin	up to 50ms (40m)	suspension fallout, bottom currents	glacimarine	Loch Lomond Stadial
K4 i	weak	weak, sub parallel	basin fill	up to 115ms (104m)	bottom currents	proximal glacimarine.	Lateglacial deglaciation
K4 ii	strong	strong, sub parallel	draped in basins		suspension fallout	distal glacimarine	

Unit K1 is widely distributed throughout the basins except on major topographic highs, and correlates with unit I1 in Upper Loch Linnhe, as shown by their identical seismic characteristics on either side of the Corran narrows on line 19 (Figs 4.3d and 4.4). It is composed of facies A and B, and is interpreted as marine sediments deposited in non-glacial conditions during the Holocene, for the reasons outlined for unit I1 in Upper Loch Linnhe.

Unit K2 is present throughout both basins. It consists of facies A in the Shuna basin, and facies B in the Kentallen basin, where there are several mounds and wedges within the unit. In places it is acoustically nearly transparent. This is characteristic of proximal glacimarine environments where high sedimentation rates result in massive depositional units (c.f. Elverhoi 1984). Unit K2 is interpreted as a glacimarine unit, probably deposited during the last glacial event in the region, the Loch Lomond Stadial. The reflectors of unit K2 slope southwards from the Corran narrows (Fig 4.5c), and also from Kingairloch, suggesting sediment sources in these two areas. Sediment supply from the Corran area is also supported by the difference in facies between the two basins; this difference may reflect the difference between sedimentation in an inner proximal zone in a glacimarine environment, characterised by mass movement of unstable sediment in high energy conditions, and sedimentation in an outer proximal zone, where deposition from suspension predominates (c.f. Syvitski and Praeg 1989, Boulton 1990). If so, both lines of evidence suggest that the Loch Linnhe ice front lay close to the northern margin of the Kentallen basin when unit K2 was deposited. The unit is thickest in the deeper parts of basins and pinches out to the north along lines 1 and 2 to form wedge shaped structures (Fig 4.5c). One of these wedge structures continues down loch as a mound (Fig 4.5b), and may represent an ice contact submarine outwash fan, analogous to those present immediately onshore at this point in the loch (Section 2.3.3, Thorp 1986). Truncated reflectors within this unit at two points in the Shuna basin suggest that it has been eroded since deposition. One of these locations is immediately adjacent to a topographic high, and the erosion may thus be due to scouring by debris flows activity and associated turbidity currents. Boulders within the same unit along lines 12 and 14 may represent iceberg rafted debris.

Unit K3 is present along line 19 (Fig 4.3c), and possibly on line 4 , both just south of the Corran narrows. It is composed of facies D, and the constructional form and irregular internal structure suggests deposition in contact with, or close to, glacier ice. The unit may be composed of submarine moraines marking glacial stillstand positions. Studies of contemporary tidewater glaciers in Alaska have shown that advance through deep water occurs on submarine moraine shoals (Post 1975, Mayo 1988), which are trundled forward at the terminus during advance and remain in the maximum position when retreat commences. Existing terrestrial evidence suggests that the Linnhe glacier reached its maximum in this basin (Thorp 1986). Hence one or both of these presumed submarine moraines may well mark the maximum of the Loch Lomond Stadial Linnhe glacier.

Unit K4 is widely distributed in the basins. It is usually possible to distinguish two sub units. The lower one is composed of facies B, infilling basins and with few weak internal reflectors, whereas the upper is composed of facies A and contains numerous strong internal reflectors. This unit is interpreted as a glacimarine unit, deposited during Lateglacial deglaciation. The surface and internal reflectors slope from north to south (Fig 4.5c), the surface being lowest in the deepest basins. This suggests sediment supply from around the Corran narrows. The characteristics of the lower subunit may reflect deposition in an ice proximal environment where high energy conditions and unconsolidated sediments result in debris reworking and resedimentation into basins. The upper subunit may represent deposition in an ice distal location where suspension fallout from a surface plume predominates. The strong internal reflectors in this lower distal glacimarine unit may reflect sporadic sedimentation and erosional events, for example associated with variable meltwater effluxes from the ice terminus and associated strong bottom current activity. These characteristics support the view that the unit represents a deglacial sequence.

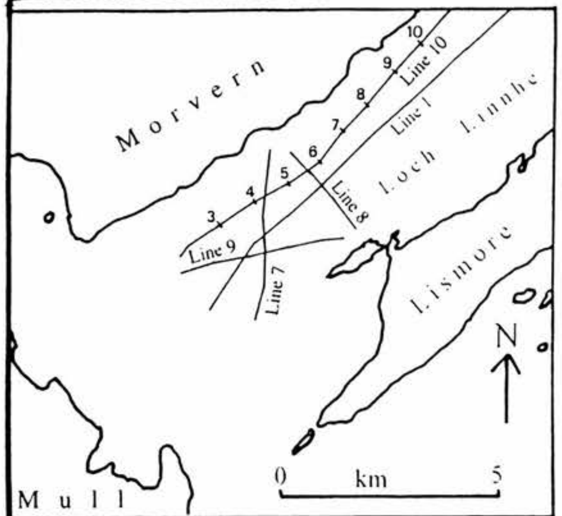
Within this unit on line 5 there are examples of facies C adjacent to the basin walls. It is suggested that these are debris flow deposits resulting from reworking of unconsolidated and unstable deposits. On line 13 unit K4 shows a complicated internal structure, possibly related to debris flow or turbidity current activity around the adjacent topographic high.

4.4.3 Lismore basin

The Lismore basin is at the southern end of Loch Linnhe (Fig 4.6a). Seismic profiles show the bedrock basin reaches a maximum depth of 280m and contains a maximum sediment thickness of 131m in the southern part of the basin. The record in the northern section of the basin suffers from acoustic blanking. Figure 4.6 shows the arrangement of units in the Lismore basin, and Table 4.3 summarises their characteristics and the interpretations of their origins. The limited distribution of the units and the limited clarity of parts of seismic records mean that these interpretations are not well constrained.

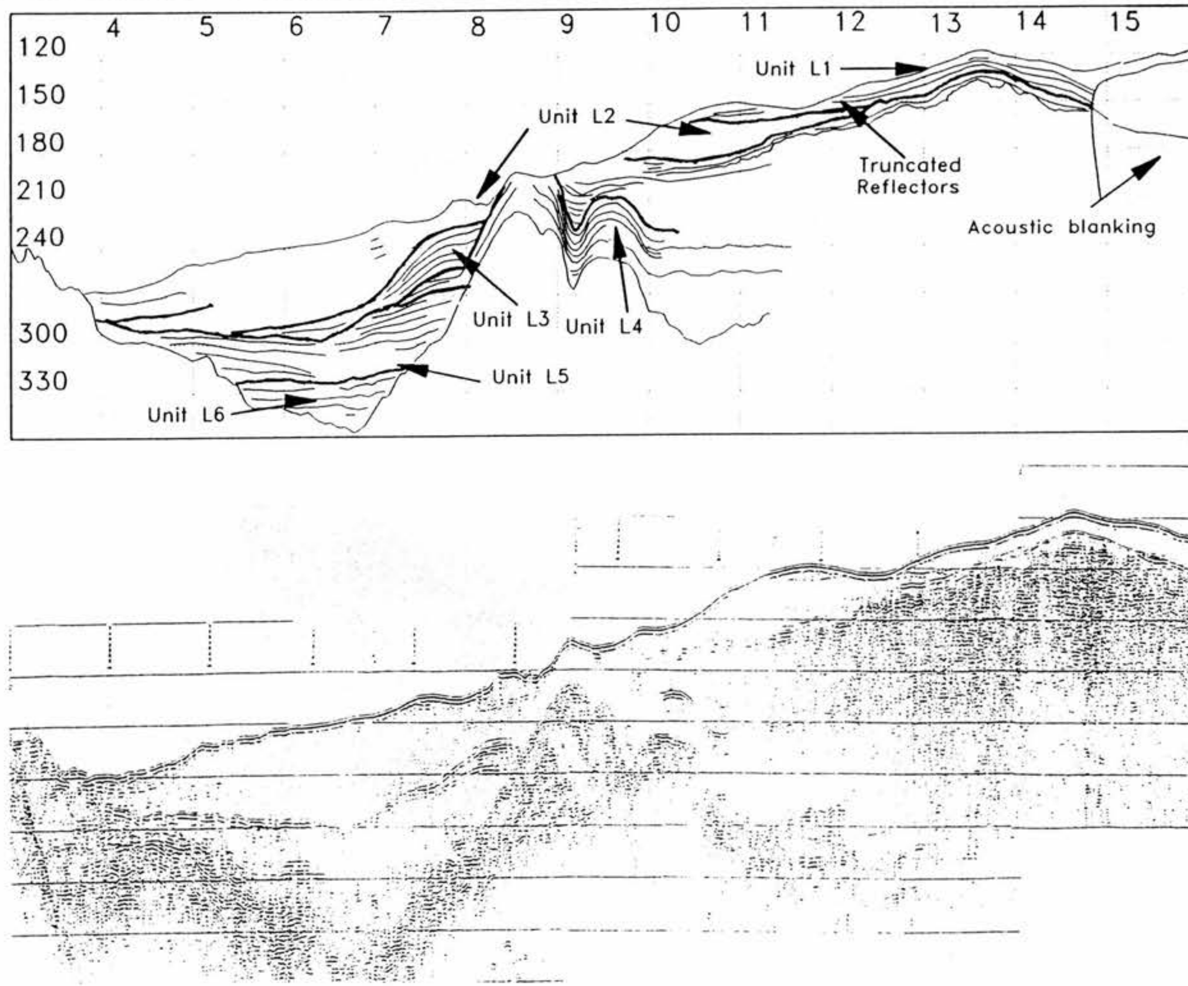
Table 4.3 Units in the Lismore basin.

Unit	Surface reflector	Internal reflectors	Geometry	Thickness	Environment of deposition	Possible age
L1	strong	strong, sub parallel, truncated	draped, wedges, mounds	up to 14ms (12m) (line 1 >40 ms?)	suspension fallout	Loch Lomond Stadial?/ Holocene?
L2	variable	weak, incoherent	infill	up to 54ms (49m)	in situ, ice contact?	Lateglacial maximum ?
L3	weak	fairly strong , sub parallel	infill, slumped	up to 39ms (35m)	bottom currents, debris flows	
L4	fairly strong	fairly strong, sub parallel	draped, slumped	up to 37ms (33m)	suspension fallout, debris flows	pre Lateglacial
L5	fairly strong	weak, sub parallel, discontinuous	draped in basins, slumped	up to 42ms (38m)	suspension fallout, debris flows	maximum?
L6	medium	medium, sub parallel	infill	up to 34ms (31m)		



a. Location of survey lines, and fixes shown in Fig 4.6b.

Fig 4.6 Seismic stratigraphy of the Lismore basin.



b. Line 10 - seismic stratigraphic units and seismic original.

Unit L1 is composed of facies A, suggesting deposition from suspension. It is present in greater thicknesses in the north of the basin than in the south, and the internal reflectors within the unit in the south are truncated, suggesting that it has been eroded since deposition. It may be very thick to the north of the basin, but these parts of the records are obscured by acoustic blanking making definite distinctions between units impossible. This may be a unit deposited during the Holocene which is presently being eroded in places, or possibly a partly reworked distal glacimarine unit dating either from the LLS or earlier.

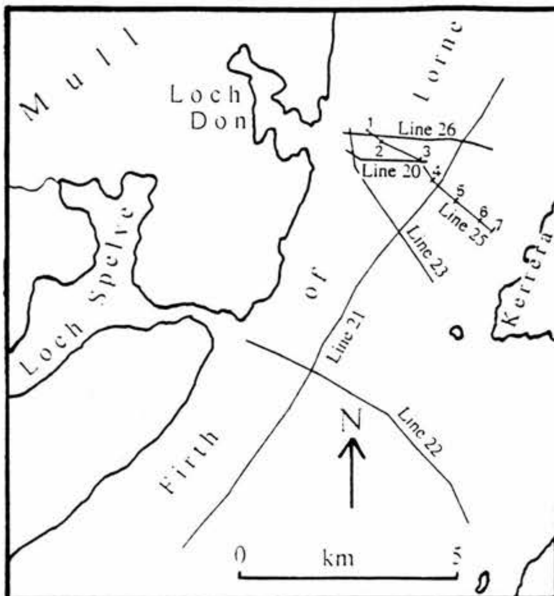
Unit L2 is widely distributed in this basin, and may be composed of facies D. On some lines it has an incoherent structure and several point reflectors, which may be reflections from boulders within a diamicton. If so, this unit may represent a glacigenic diamicton deposited close to, or in contact with glacier ice. The last period in which ice extended into the Lismore basin was during the Dimlington Stadial, and so this unit may date from this event. The unit has an irregular, erosional lower boundary, suggesting erosion prior to deposition of this unit, and it also contains truncated internal reflectors indicating post-depositional erosion.

Units L3 - L6 are each of limited distribution, units L5 and L6 being only present in the deepest part of the basin. They consist of facies B,A and C, and may represent glacimarine units. The limited distribution of each, and the numerous truncated reflectors in units L3-L5 suggest that these units were deposited in a part of the sea loch where high energy conditions, and an erosional oceanographic regime have prevailed on numerous occasions. The very limited distribution of units L3 and L4 and the erosional upper surfaces of units L4 and L5 suggest that there has been a major erosive event in this basin, possibly between the deposition of units L3 and L4. This might be associated with ice advance prior to the Dimlington Stadial.

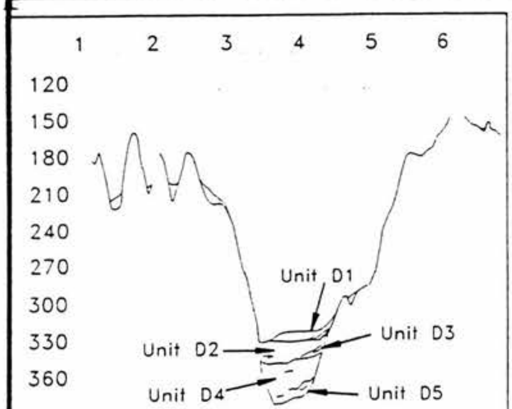
The sea bed reflectors on lines 8, and possibly lines 9 and 10, show point reflectors which may indicate a boulder lag due to erosion of fine sediment by bottom currents. Several lines of evidence thus suggest that this basin has been subject to erosional conditions on numerous occasions during the Late Quaternary.

4.4.4 Don basin

The seismic units in the area east of the Isle of Mull are thinner than those in Loch Linnhe, and sediment is largely confined to the deepest basins. Thick sediments may occur to the north of the Don basin, but the records are obscured by acoustic blanking. The surface reflector north of the Don basin is smooth and wavy, suggesting sediment bedforms produced by bottom current activity.



a. Location of survey lines, and fixes shown in Fig 4.7b.



b. Line 25 - seismic stratigraphic units and seismic original.

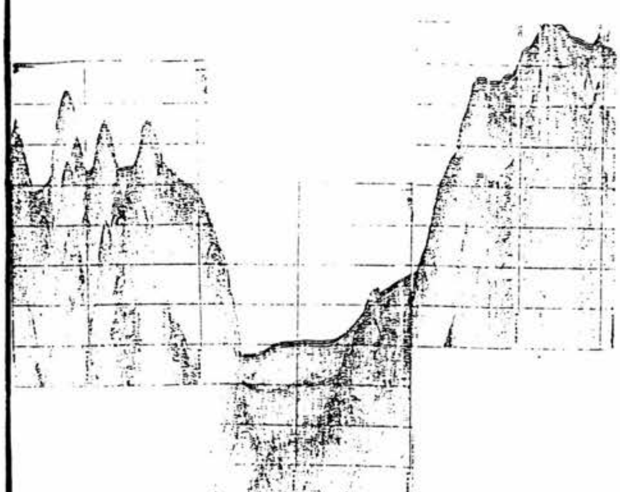


Fig 4.7 Seismic stratigraphy of the Don basin.

The Don basin is the deepest basin in the area surveyed. The bedrock basin is 325m deep and contains sediments to a maximum thickness of 79m. These sediments have been divided into 5 units (Fig 4.7). These units are recognised on lines 21 and 25 which intersect in the Don basin. They are tentatively extended to the adjacent parts of lines 20 and 26, although the locations of the lines surveyed do not prove that the basin in these latter lines is a continuation of the Don basin. The limited distribution of the units, and the lack of clarity of the seismic records mean that the following interpretations are provisional.

Unit D1 is locally present, up to 10m thick and composed of facies C debris flow deposits.
Unit D2 is present throughout the basin, is up to 22m thick, has a strong, smooth surface reflector, and probably consists of facies B. This unit may be a glacimarine unit dating from the LLS or Lateglacial deglaciation.
Unit D3 is a second local facies C debris flow unit, up to 6m thick.
Unit D4 is present throughout the basin, has a strong, smooth surface reflector and is composed of facies B. It is up to 36m thick. This may be a glacimarine unit, perhaps deposited during the Devensian glaciation.
Unit D5 is locally present at the base of the Don basin and up to 12m thick. It has a strong surface reflector and probably the characteristics of facies C. It probably represents slumped material.
Elsewhere, several other small basins contain infills of acoustically laminated facies (Fig 4.7, fixes 1-3). At the mouth of Loch Don there are seismic facies with internal reflectors that are occasionally parallel, but mostly incoherent, which prohibits identification of the seismic facies.

4.4.5 Summary

Contrasts in the distribution of seismic stratigraphic units in Loch Linnhe and the Firth of Lorne clearly mark the LLS limit, as illustrated schematically in Fig 4.8. Inside the limit possible LLS and older units are confined to the deepest part of the Inverscaddle basin. Over most of the seabed inside the limit only unit A is present, and several lines of evidence suggest that this consists of Holocene sediment. There are clear ice marginal features in the Kentallen basin marking the terminus of the LLS Linnhe glacier. In addition, in both the Kentallen and Shuna basins there are thick and widespread seismic stratigraphic units of LLS and older ages overlain by the Holocene unit. In the Lismore and Don basins down-fjord of the LLS ice limit there are probably thin or few Holocene and LLS units, and the deep basins contain units of a Late Devensian and possibly earlier age.

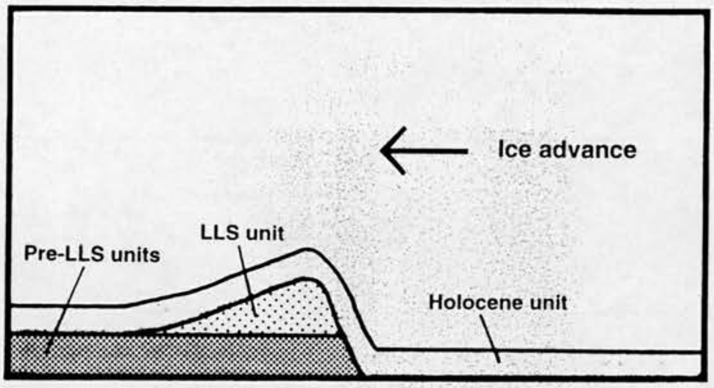


Fig 4.8 Schematic illustration of submarine sediment distribution around the ice limit in Loch Linnhe.

4.5 Discussion

4.5.1 Glacial and marine processes in Loch Linnhe

Comparison of the distribution of seismic stratigraphic units in the different basins with the established chronology of Quaternary and Holocene environments allows inferences to made about glacial and marine processes at different times in Loch Linnhe. The seismic evidence may be synthesised to construct a provisional, tentative chronology of environmental conditions in Loch Linnhe, shown in Table 4.4.

A consideration of contemporary marine circulation patterns explains the distribution of the Holocene unit, and partly explains the distribution of older units in Loch Linnhe. Under contemporary oceanographic conditions there are strong tidal currents over the Don and Lismore basins (Admiralty charts 2379, 2387). Further up Loch Linnhe the marine environment is more sheltered and of lower energy, with the exception of tidal currents through the narrows (Admiralty chart 2380). These conditions explain the distribution of the Holocene units, which are located only in areas where there are no strong tidal currents. In addition, the numerous erosional unconformities in the outer basins, and the limited distribution of sediment suggest that these areas have been subject to relatively strong tidal scour at many times during the late Quaternary (c.f. Syvitski 1989, Davies et al. 1984). Fjord circulation patterns may thus have often been similar to those of today. Due to the repeated episodes of erosion the stratigraphic record of events in these basins is unlikely to be complete. If contemporary fjord circulation patterns were the only control on the distribution of sedimentary units, the succession of older units present in the deepest parts of the outer basins

Table 4.4 Provisional environmental chronology for Loch Linnhe and the Firth of Lorne.

Holocene.	In the inner loch, strong tidal currents prevent sedimentation in shallows and narrows, but elsewhere sediment can accumulate by suspension fallout from freshwater surface plumes originating from river mouths, and this sediment is being reworked by bottom currents. The thickness of unit A suggests Holocene sedimentation rates of up to 2.4cm a^{-1}. Contemporary strong tidal currents in the outer basins probably prohibit the accumulation of Holocene deposits.
Loch Lomond Stadial.	Glacimarine deposits of relatively well bedded materials with occasional iceberg rafted debris of larger clast size accumulated mainly in the Kentallen and Shuna basins. Sediment reworking by bottom currents and debris flows was widespread in these basins. A possible accumulation of till and/or an ice contact slope south of Corran may mark the maximum of the Loch Lomond Stadial glacier. As there may be no equivalent units up-fjord, it is likely that little debris was deposited during ice retreat. The basins in the outer loch were probably affected by strong tidal currents, as they are today, which prevented the accumulation of LLS distal glacimarine sediments.
Late Devensian deglaciation.	Fairly thick glacimarine deposits accumulated in the Kentallen and Shuna basins, possibly during a stillstand at Corran narrows. There may have been some deposition in the deepest outer basins.
Late Devensian ice maximum.	Erosion of the inner lochs removed any pre-existing sediments. Till may have been deposited in the deepest basins in the outer lochs.
Previous glaciation.	Glacimarine sediments deposited in deepest basins in the Firth of Lorne?

should also be present in the upper basins of Loch Linnhe. The lack of these older units suggests that there may have been major erosive events in the inner loch, which removed older units from these relatively shallow basins. The spatial distribution of seismic units of different ages thus suggests that there has been an ice advance to the Kentallen basin (Fig 4.8).

The lack of LLS and older seismic stratigraphic units in the Inverscaddle basin suggests that the LLS Linnhe glacier may have eroded most of the older sediment units in this basin (c.f. Carlson et al. 1983, Molnia et al. 1984). Most of this eroded sediment was probably deposited as thick glacimarine units in the Kentallen and Shuna basins, which were the immediate proglacial basins (c.f. Powell 1981, 1983, Elverhoi 1984, Syvitski 1989). The restricted volumes of LLS glacimarine units in the outer basins is explained by their remoteness from LLS ice termini, and a likely high energy fjord circulation pattern here.

Inferences may also be made about pre-LLS glacial action. In the Kentallen and Shuna basins the basal stratigraphic unit shows the characteristics of a deglacial glacimarine sequence, and most probably dates from Late Devensian deglaciation. The internal reflectors of this unit suggest that at least part of it may have been deposited during a glacial retreat stillstand around the Corran narrows. The lack of seismic evidence for older sediments in the Inverscaddle, Kentallen and Shuna basins suggests that Late Devensian ice advance removed all pre-existing units from these basins. The presence of a possible Late Devensian till and maybe even older units in the deepest outer basins, suggests that ice during the lateglacial maximum overrode pre-existing sediments in these locations.

The distribution of seismic stratigraphic units in the different basins is thus explained by a consideration of the extent and erosive activity of ice during different glaciations, and fjord circulation patterns in Loch Linnhe. In particular, the arrangement of the seismic stratigraphic units indicates a lateglacial ice advance to beyond the Corran narrows. The evidence in Loch Linnhe and the Firth of Lorne supports the view that deposition through a glacial cycle is episodic, being concentrated at locations which change position through the cycle (Powell and Molnia 1989). During a glacial cycle, within fjords, glacimarine deposits may be concentrated in the proglacial marine basins associated with ice maxima and retreat stillstands.

Facies C, which is interpreted as resulting from slumping and debris flow activity, is locally present in each basin. In the Inverscaddle and Kentallen basin it occurs in units I1 and K1 respectively, and indicates movement from the Corran narrows to the basins both north and south. In the Kentallen and Shona basins slumped facies indicate flows from topographic highs in all units, and also from the Corran narrows in unit K4. Facies C is also present in units L3 and L5 in the Lismore basin indicating movement from the topographic high at the north of the basin, and in the Don basin, indicating debris flows from numerous submarine high points, in particular the mouth of Loch Don. The prevalence of these deposits can be explained by the high relief fjord bottom topography, and a consideration of likely triggers for debris flow events.

Very strong tidal currents through the Corran narrows throughout the late Quaternary are likely to have scoured debris from this topographic high to the basins on either side. In addition, episodes of deep water renewal may lead to the transport of fines from surficial sediments landwards of the sill to deeper parts of the basin (Gade and Edwards 1980, Syvitski 1989).

Numerous authors have concluded that slumps and debris flows are widespread following deglaciation (e.g. Powell 1981, 1991, Aarseth et al. 1989, Elverhoi 1984, Carlson et al. 1983), mainly for three reasons. Firstly, large volumes of unconsolidated and unsorted materials are deposited by a glaciers in unstable positions (Syvitski 1989). Secondly, slumps and flows can be triggered by sea level rise. This increases the pore water pressure in the sediment which decreases internal friction and cohesion, making slumping more likely. The Lateglacial period was one of rapidly rising eustatic sea levels, which would later have been exceeded by isostatic recovery of the land surface (Lambeck 1991). A further trigger for slumping events is seismicity (Carlson 1989). It is probable that the Lateglacial was a period of relatively high seismic activity (Holmes 1984, Ringrose 1987, Sissons and Cornish 1982). This is because crustal movements would have been suppressed during the preceding glaciation by the large mass of ice sitting over the crust, and because the deglaciation entailed a large amount of isostatic readjustment in the continental crust.

4.5.2 Comparison with other offshore evidence.

There are interesting differences and similarities between the pattern discussed here and that described by Boulton et al. (1981) concerning the distribution of seismic stratigraphic units of different ages in the sea lochs and sounds around Arisaig. In both locations, thick sequences of Devensian deglacial deposits and thinner units of other pre-LLS deposits are rare inside the LLS limits, but are present in sheltered basins outside the limits, suggesting that LLS ice advance removed most pre-existing sediments.

The main difference in the distribution of sediment between the two areas is in the location of proglacial LLS deposits. In Lochs Nevis and Ailort the LLS ice maximum is marked by submarine terminal moraines at the mouths of these fjords. Unlike the situation in Loch Linnhe, there are no thick sequences of glacimarine sediment immediately outside the LLS ice limits, but there is seismic and borehole evidence suggesting distal LLS glacimarine sedimentation in basins further west (Davies et al. 1984, Fyfe et al. 1993, Stoker et al. 1989,1993). Most of these LLS sediments are part of a larger seismic stratigraphic unit which is 10-30m thick on the continental shelf around the inner and outer Hebrides, but is thickest around the inner Hebrides, locally exceeding 300m in deep basins. It is likely that this lack of LLS proglacial sedimentation close to the palaeo-ice fronts at the mouths of Lochs Nevis and Ailort reflects oceanographic conditions. Under contemporary conditions strong tidal currents flow parallel to the coastline, and these may entrain sea bed sediments as water depths in the inner sounds are commonly shallow. It is likely that similar tidal currents prevailed during the LLS, when sea-levels were within a few metres of present levels.

By contrast, in Loch Linnhe, the LLS glacier terminated within the fjord, and bottom currents within fjords are relatively weak (Boulton 1990). Proglacial sediment thus accumulated in the deep proglacial basin. It is possible that additional distal glacimarine sediments from the LLS Linnhe glacier accumulated further south, as there is evidence of iceberg rafted debris within a sediment unit spanning the Lateglacial in the approaches to the Firth of Lorne (Davies et al. 1984, Fyfe et al. 1993). However, there is as yet no evidence to suggest to what extent these distal sediments relate to the Linnhe glacier and/or LLS glaciers which terminated at the mouths of Lochs Creran, Etive, Spelve and Don.

In Lochs Linnhe, Nevis and Ailort, there is little evidence of LLS deglacial glacimarine units inside the LLS ice limits, but there are possible stillstand morainic deposits at sills and

narrows in Lochs Nevis (Kylesknoydart) and Ailort (Roshven) which both correlate with proglacial outwash fans on land. Evidence in each loch thus suggests deglaciation may have been punctuated by stillstands at narrows and shallows.
The distribution of Holocene sediment is similar in both areas, with deposition being concentrated in the sheltered sea lochs and deep basins in the sounds.

This contrast in the distribution of LLS proglacial sediment supports patterns observed elsewhere. In Alaska, when glaciers terminate within fjords sediment is trapped in the fjord basins, and the continental shelf is starved of sediment. When glaciers advance to the fjord mouths the fjord basin infills are eroded and transported by the glacier and ocean currents to the continental shelf and slope (Powell 1991). The location of the LLS ice maxima within different fjord systems in Western Scotland may thus have placed important controls on the locations of the main centres of deposition during the stadial.

4.6 Summary

1. There is a clear seismic stratigraphy in much of Loch Linnhe showing deep bedrock basins with thick sediment infills.
2. The geometry, internal characteristics and location of LLS seismic stratigraphic units in Loch Linnhe and the Firth of Lorne indicates that the Linnhe glacier terminated in the northern half of the Kentallen basin. This glacier may have eroded much of any pre-existing sediment in the Inverscaddle basin during advance, and deposited little during retreat through deep water.
3. The evidence suggests that Holocene sediments have accumulated in the sheltered inner basins, but not in the outer basins which experience strong tidal currents.
4. The characteristics of a proposed Devensian deglaciation glacimarine unit suggest that there may have been a retreat stillstand around the Corran narrows.
5. Possible older sediment units are confined to the outer basins in Loch Linnhe and the Firth of Lorne. These show evidence of erosion, which may be linked to Devensian ice advance and / or strong tidal currents on numerous occasions during the Quaternary.
6. The spatial distribution of LLS glacimarine units in Western Lochaber suggests that deposition was concentrated in locations associated with the ice maxima and retreat stillstands.

Chapter 5 - Synthesis of the field evidence; The reconstructed ice cap

5.1 The reconstructed ice cap

This chapter synthesises the field evidence described in Chapters 2, 3 and 4 and reconstructs the morphology and retreat pattern of the LLS ice cap in Western Lochaber.

The reconstructed ice cap is shown in Fig 5.1. At higher altitudes in the upper reaches it is based on trimline altitudes and ice flow direction indicators such as striations and erratic evidence. In the lower reaches it is based on geomorphological evidence of ice limits, ice flow direction indicators and seismic survey results.

The surface contours are drawn according to the guidelines given by Sissons (1974b), with concave contours in the upper parts of the glaciers, and convex contours in the lower sections of land based glaciers. In the lower reaches of marine terminating glaciers, however, contours were kept concave, as these glaciers have extensional rather than compressional flow at the grounding line and this draws down the glacier surface to a concave cross profile (Powell 1991). Ice contours were plotted at 90^{0} to the ice flow direction indicated by flutings, striations, crescentic gouges and roche moutonnées.

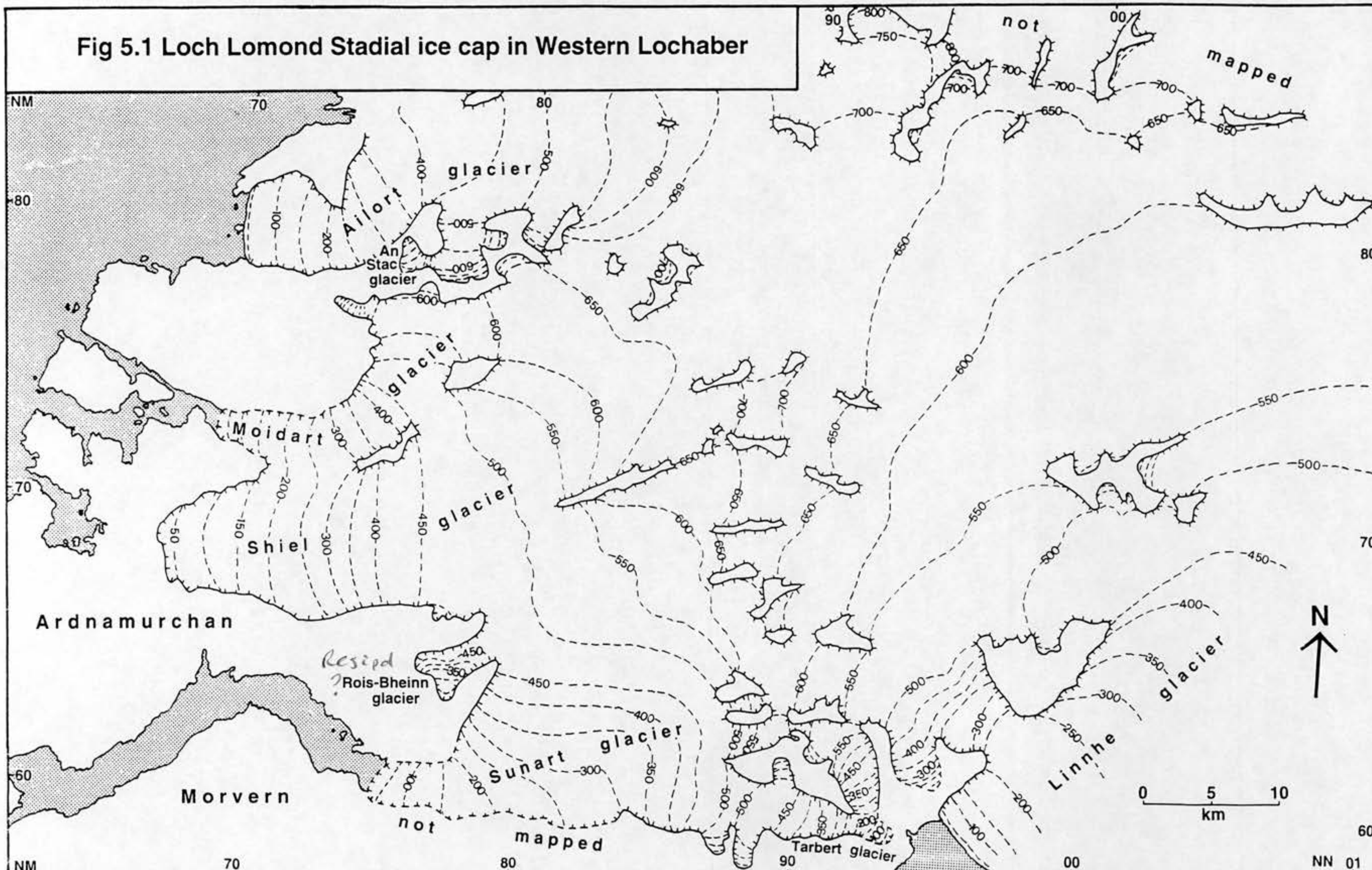


Fig 5.1 Loch Lomond Stadial ice cap in Western Lochaber

Fig 5.1 shows a large, fairly flat ice field with a surface altitude of 600-650m in the centre and north east of the area, with outlet glaciers that flow down the sea loch troughs to the west and south. Some of the ice from the north east of the area probably flowed eastwards as well as southwards out of the study area. In general, ice flowed down the major troughs, but in some places it was thick enough to cross low cols and ridges linking adjacent troughs. The ice field was fed from ice in the central mountain belt in Ardgour, and from the higher mountains to the north and north west. It is assumed that all the tidewater glaciers were grounded, since there are no contemporary examples of temperate tidewater glaciers with floating termini, and since it is thought that the tensile strength of temperate ice is too low for it to advance by floating (Powell 1988).

5.2 Evidence for terminal limits

In some troughs there is direct evidence for the terminal positions of the outlet glaciers, such as moraines and ice contact slopes. Elsewhere, indirect evidence, such as the inner limit of high, lateglacial raised beaches, is used to constrain the former ice maximum to particular sections of troughs.

5.2.1 Direct evidence for ice limits

The clearest terminal limit is at the snout of the former Shiel glacier. The southern section of the ice terminus is recorded by a terminal moraine at 50m OD on the rock bar west of Acharacle, and the northern section by an ice contact slope on the outwash plain (Fig 2.15). The interpretation that these features mark the maximum extent of the ice is supported by the sheer volume of material deposited immediately in front of the paleao-ice margin as it is unlikely that such a large volume would be deposited during a retreat stage. This proglacial outwash grades away from the maximum limits both at the ice contact slope (Shennan et al. 1994b, Dawson 1994), and from a meltwater channel cutting through the rock bar on which the moraine lies (Wain-Hobson 1981), and continues for 3.5km.

In the Linnhe trough, terrestrial and submarine evidence both indicate an ice maximum south of the Corran narrows. Thorp (1984, 1986) placed the maximum limit at the ice contact slope of the southmost of a series of four ice marginal outwash fans. The inner three fans he interpreted as retreat stillstand deposits (Fig 2.13). Evidence presented here from the sea bed and the west shore of the loch support this limit for several reasons. Firstly, the results of the seismic survey showed that the thickest spreads of LLS glacimarine deposits lie in the basin south of the Corran narrows. Glacimarine sediments are normally concentrated in the proglacial submarine basin (Powell 1981, 1983). Secondly, the arrangement of internal reflectors in the LLS seismic stratigraphic unit also suggests a former ice front in this basin. These reflectors suggest high energy and ice contact environments of deposition. The limit may be marked by the mounds of glacial till found just south of the Corran narrows, and /or the ice contact slope within the LLS glacimarine unit (Section 4.7.2), which corresponds exactly with the location of the outwash fans onshore. Thirdly, the concentration of outwash and raised beach deposits around the shores of the loch in this area can best be explained by the former presence of an ice front.

The reconstructed limit in Loch Ailort is at the mouth of the loch and corresponds with the submarine moraine shoal found by Boulton et al. (1981) in a seismic survey of the surrounding waters. The limit is also marked by contrasts in glacial deposits and shoreline erosion inside and outside the submarine moraine. For example, there are occasional terrestrial till deposits inside the limit at Roshven and no clear till deposits outside the limit. Comparison of the shores and islands shows both that those outside the limit are more frost shattered and degraded than those inside, and that a raised cliffline is much clearer outside the limit (e.g. on Eilean nan Gobhar) than inside. One explanation for this pattern is that the shores outside have been subject to a longer period of weathering and erosion than those inside. The lateral ice limit east of Roshven may be marked by a patchy drift limit on slopes on the southern side of the loch at 250-300m.

The terminal limit of the outlet corrie glacier south of Beinn Resipole is marked by a series of morainic mounds on a bedrock bar in the glen. The western lateral margin is defined by a clear limit to drift, small moraines and boulders.

The glacier in east Glen Tarbert is likely to have advanced to a position close to the mouth of the glen on Loch Linnhe. This is suggested by both the thinning of drift cover and the large volume of proglacial outwash at Inversanda in east Glen Tarbert.

5.2.2 Indirect evidence for ice limits

The locations of 'high' raised marine features (Section 2.3.7, Fig 2.21), provide important additional constraints on ice limits. As a glacial advance is likely to erode beach deposits from coasts, the presence of these beaches shows that the sites concerned were not covered by LLS ice.

The ice limit in the Moidart trough is poorly constrained as there is no terminal or stillstand evidence at any point in lower Glen Moidart or around the loch. The limit has been tentatively placed in the inner half of the sea loch. This is because the presence of high raised beaches at

20-30m and possible marine sediments at 9.33m (Section 2.3.7) suggest that ice did not extend to the outer half of the sea loch.
There is no clear evidence for the extent of the LLS glacier in Loch Sunart. The presence of high raised beaches mapped by the BGS (sheet 52, drift edition) in the outer part of the loch suggest the glacier did not extend as far as Glencripesdale. Features indicating palaeosea-levels of 9-11m OD at Resipole may mean the glacier did not extend this far, although a precise age cannot be inferred from this altitude as there is no established sea-level curve for this loch. Here, it is suggested that the limit may lie close to the Sunart narrows opposite Laudale, as there is a concentration of raised beaches below 10m O.D. at this point in the loch which might represent raised and reworked glacial stillstand deposits.

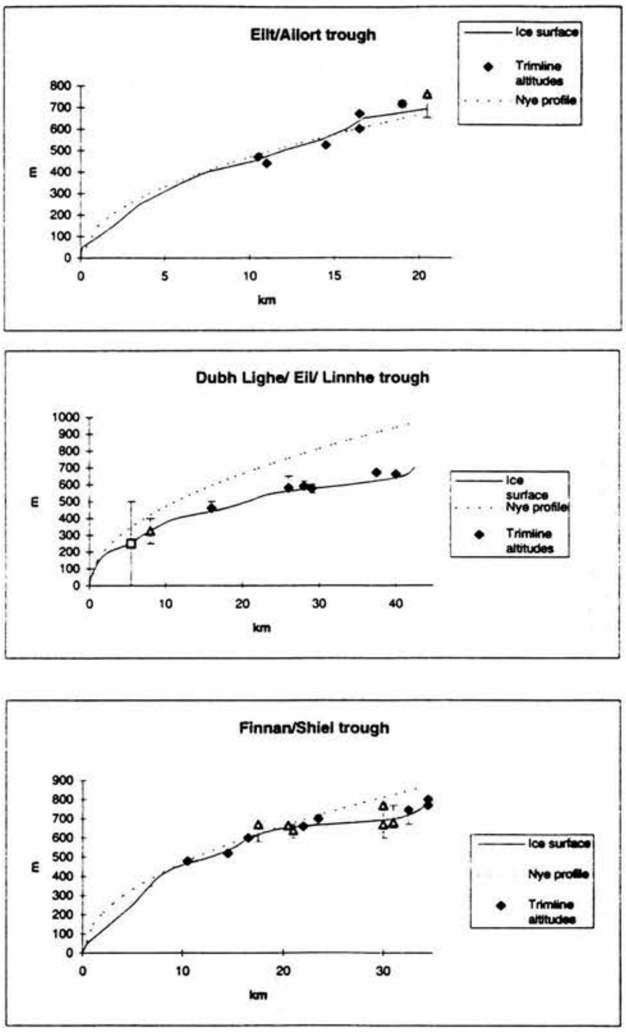
The tentative limits in these two troughs where there is just indirect evidence for ice limits are consistent with the limits in the other troughs which contain clear, direct evidence as they indicate palaeoglaciers of comparable lengths.

5.2.3 Validity of the reconstruction

Terrestrial geomorphic evidence, trimline evidence and marine seismic evidence are independent lines of evidence, yet all agree on the reconstructed ice limits in Western Lochaber. This coincidence of the different palaeo-ice cap indicators suggests that the reconstruction of LLS ice cover is reliable. Several arguments support this conclusion.

1. The trimline evidence for the ice surface in the upper reaches of the former glaciers is widespread and consistent. The way the trimline values and limits in troughs cross-check across glens and conform between adjacent glens both supports their validity (Fig 3.4, Section 3.6.1)

2. The reconstructed ice surfaces suggested by trimline altitudes are entirely consistent with the generally down-trough ice flow directions suggested by the ice flow directional indicators discussed in Chapter 2. These indicators include striations, roche moutonnées, glacial flutes and erratic transport pathways. Furthermore, independent evidence of ice flow directions in this area in the form of striations and roche moutonnées mapped by the British Geological Survey (Sheets 52, 61, 62, drift editions, Sheet 53) are all in accord with the ice flow directions shown in the ice cap reconstruction presented here.



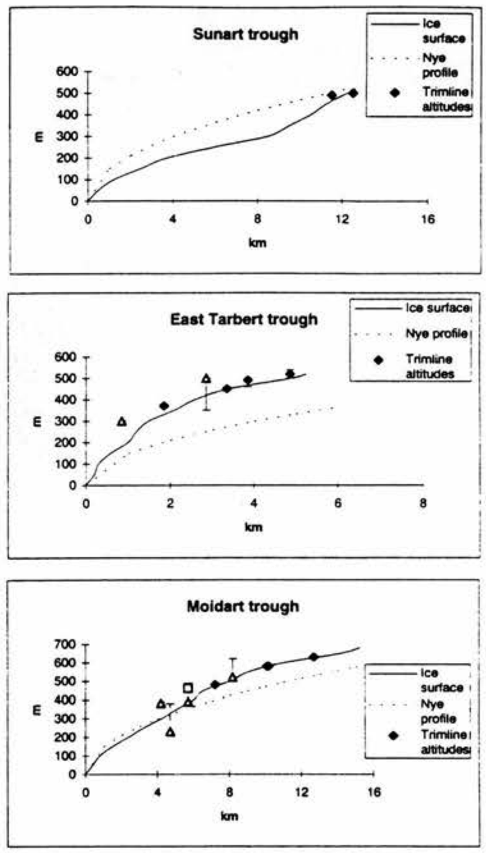
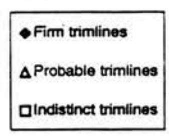


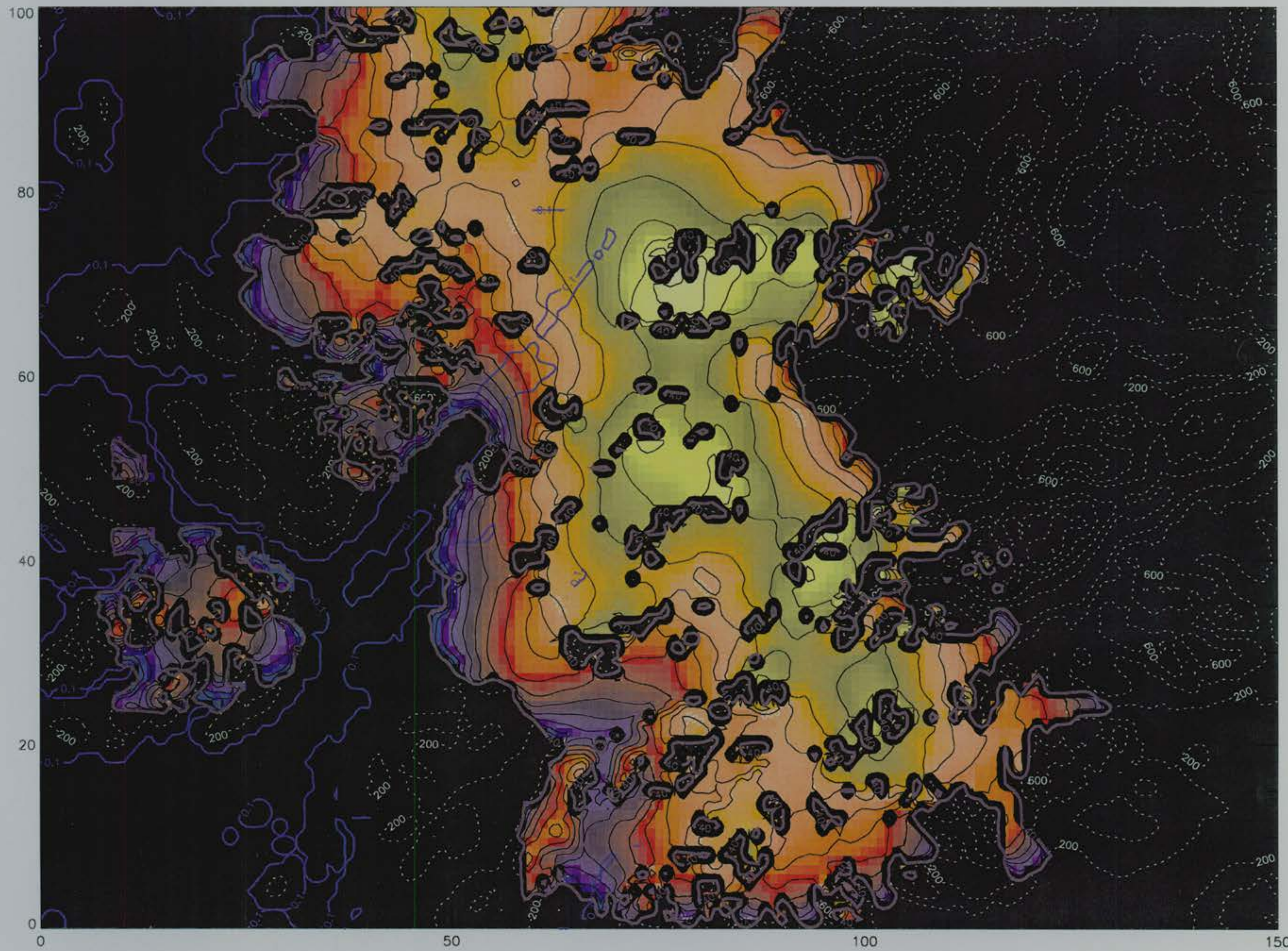
Fig 5.2 Glacier surface profiles and theoretical ice profiles.



3. The ice surface profiles reconstructed from field evidence are comparable with profiles derived from glaciological theory. This is shown in Fig 5.2 where a theoretical Nye profile drawn from the terminal limits in the Shiel and Ailort troughs coincides with the points marked by trimline altitudes, and conforms closely with the reconstructed ice surface profiles. The surface profile of the reconstructed Linnhe glacier is lower than the theoretical profile, which is common in outlet glaciers (Sugden 1977). The East Tarbert glacier has a higher surface profile than that derived from glacial theory. This may be partly due to the steep slope of the glen floor in part of this tough, or may suggest that this glacier advanced further towards Loch Linnhe than the position shown in Fig 5.1.
In the Moidart and Sunart troughs, where there is no clear geomorphological evidence for ice maxima, the limits suggested here were selected to allow reasonable ice surface gradients to be reconstructed from the closest trimlines to these limits, within the constraints of raised beach evidence. Fig 5.2 shows that the reconstruction in the Moidart trough fits well with that predicted on the basis of glaciological theory, and the surface of the Sunart glacier is in accord with the lower profiles possible in outlet glaciers.

4. The reconstructed ice cap is of similar dimensions to an ice cap produced by numerical ice sheet modelling experiments (Hubbard, in prep.). These experiments use a high resolution coupled ice sheet - mass balance model comprising of an ice accumulation and ablation model coupled to an ice flow model. It operates on a 1km topographic grid and is driven by imposed deviations of temperature and precipitation from present day distributions. These deviations should be regarded as qualitative indicators of climate change, rather than absolute values, due to a lack of calibration in the mass balance model. The model is able to reproduce the maximum limits of the West Highland LLS ice cap reconstructed elsewhere from field evidence. A variety of different climate scenarios, each using different sets of boundary conditions and run times, result in similar outer ice limits in Lochaber, suggesting that ice advances to similar positions, but at different rates under different climatic scenarios. The model has been run with and without taking iceberg calving termini into account. In both scenarios the outer limits at the west coast of Lochaber are close to those reconstructed here, but ice morphology in Lochaber is different. When calving is not included, the Ben Nevis area dominates the West Highland ice cap so that ice flows radially from this centre of accumulation. Ice flow is thus westwards across the north of Western Lochaber, south-westwards across central Ardgour, and southwards to a terminus by the Island of Lismore, far down Loch Linnhe.

Fig 5.3 (overleaf) LLS ice cover generated by a high resolution numerical ice sheet model. From Hubbard, in prep.


100
80
60
40
20
0
0
50
100
150
600
200
0.1
40

When the effects of calving are included, the Ben Nevis area remains dominant when July temperature depressions of up to -8°C are modelled. At temperature depressions greater than this, additional independent centres of ice dispersal are produced around the mountains in Western Lochaber, and ice in Loch Linnhe terminates close to the limits reconstructed here. Fig 5.3 shows the ice cap produced after 300 simulated years using an imposed mean July temperature depression of -8.5°C, and approximately 75% of present precipitation in Western Lochaber. This corresponds closely with the maximum ice limits and surface altitudes of the reconstructed ice cap in Western Lochaber. The Shiel glacier produced by the model is larger than that reconstructed here, which may reflect the fact that the model does not incorporate calving into Loch Shiel during advance. Other climatic scenarios using temperature depressions of -5°C to -9°C, but with longer and shorter run times respectively, produce similar outer ice limits. This good fit of the field evidence with the dimensions of an ice mass produced using glaciological principles supports the validity of the reconstruction presented here.

5. The ice surface reconstructed here corresponds well with those reconstructed by Thorp (1986) and by Sissons (1979a) for the region immediately to the east of Western Lochaber. Where this and Thorp's reconstructions join along Loch Linnhe, ice surface contours differ by a maximum of 50m at any point (see Fig 6.5).
The extent of LLS ice reconstructed in this study is, however, considerably greater than that envisaged by Wain-Hobson (1981). His map of the LLS glaciers in Morvern, southern Ardgour and Sunart shows only small valley glaciers that had not coalesced to form an ice cap (Fig 1.6). Much of this work was based on the mapping of vertical drift limits, which he interpreted as ice limits. These are reinterpreted in this study as being topographically controlled, as they are mostly located at the breaks of slope beneath the steeper upper walls of the glens, and the trimline evidence discussed here suggests that many of the vertical ice limits mapped by Wain-Hobson are underestimated.

5.3 Evidence for the age of the reconstructed ice cap

Radiocarbon dates, the stratigraphy of enclosed basins, and the altitude of raised marine deposits all suggest that the reconstructed ice cap formed during the LLS. In addition, theoretical considerations concerning the development and characteristics of trimlines (Chapter 3) support an LLS age for the ice cap.

5.3.1 Lateglacial litho- and pollen stratigraphy

The litho- and bio-stratigraphic record in enclosed depositional basins provides evidence about environmental conditions in the basin catchment since deposition commenced. Investigations elsewhere in Scotland have shown that sites outside LLS ice limits may contain a tripartite Lateglacial stratigraphy. This consists of basal minerogenic sediment, often an inorganic lacustrine silty clay, overlain by a sequence of organic, minerogenic and further organic sediments. The two organic layers can be shown through radiocarbon dating or pollen stratigraphies to date to the Lateglacial Interstadial, and the Holocene, respectively (Sissons et al. 1973). The minerogenic layer between these organic units represents sedimentation during the harsh climatic conditions prevailing during the LLS.

In contrast, sites inside the LLS ice limits contain basal minerogenic sediments overlain only by postglacial organic sediments. However, not all sites outside LLS limits contain Interstadial organic sediments and not all sites inside the limits have early postglacial sediments, as local conditions may not have been suitable for sedimentation immediately after deglaciation. The absence of the tripartite pattern is thus *not* positive evidence for the site being glaciated during the LLS, although the presence of Interstadial sediments *does* indicate that the site was not covered by LLS ice (Lowe and Walker 1984). Consistent contrasts in basin stratigraphy inside and outside inferred former ice limits can support hypotheses of their age (Gray 1975, Tipping 1988, 1989).

Early Holocene sediments also contain a characteristic pollen stratigraphy showing temporal vegetation changes from pioneer to arboreal which have been shown to be regionally consistent throughout Western Scotland (Lowe and Walker 1991). These are sufficiently well established to allow different ages within the early Holocene to be inferred from the pollen assemblages present, as there was a regionally consistent succession .

Three previously investigated sites outside the reconstructed limits around the lower Shiel trough contain tripartite Lateglacial stratigraphies. The locations of these sites are shown in Fig 5.4. Wain-Hobson (1981) cored two basins between the Sunart trough at Salen and Shiel trough at Acharacle, finding the tripartite litho-stratigraphy at NM 692 646 and NM 693 654. provide
Pollen investigations and a radiocarbon date confirmed that the basal organic sediments at the here
latter site are of Lateglacial Interstadial age. Shennan and Walker (1994) have found the tripartite litho-stratigraphy in a basin seawards of Kentra Moss (NM 633 707). Diatom analyses suggest the basal sediments here are of Lateglacial Interstadial age, and further pollen and radiocarbon analyses are in progress. These results show that these sites have not been glaciated since the Lateglacial Interstadial.

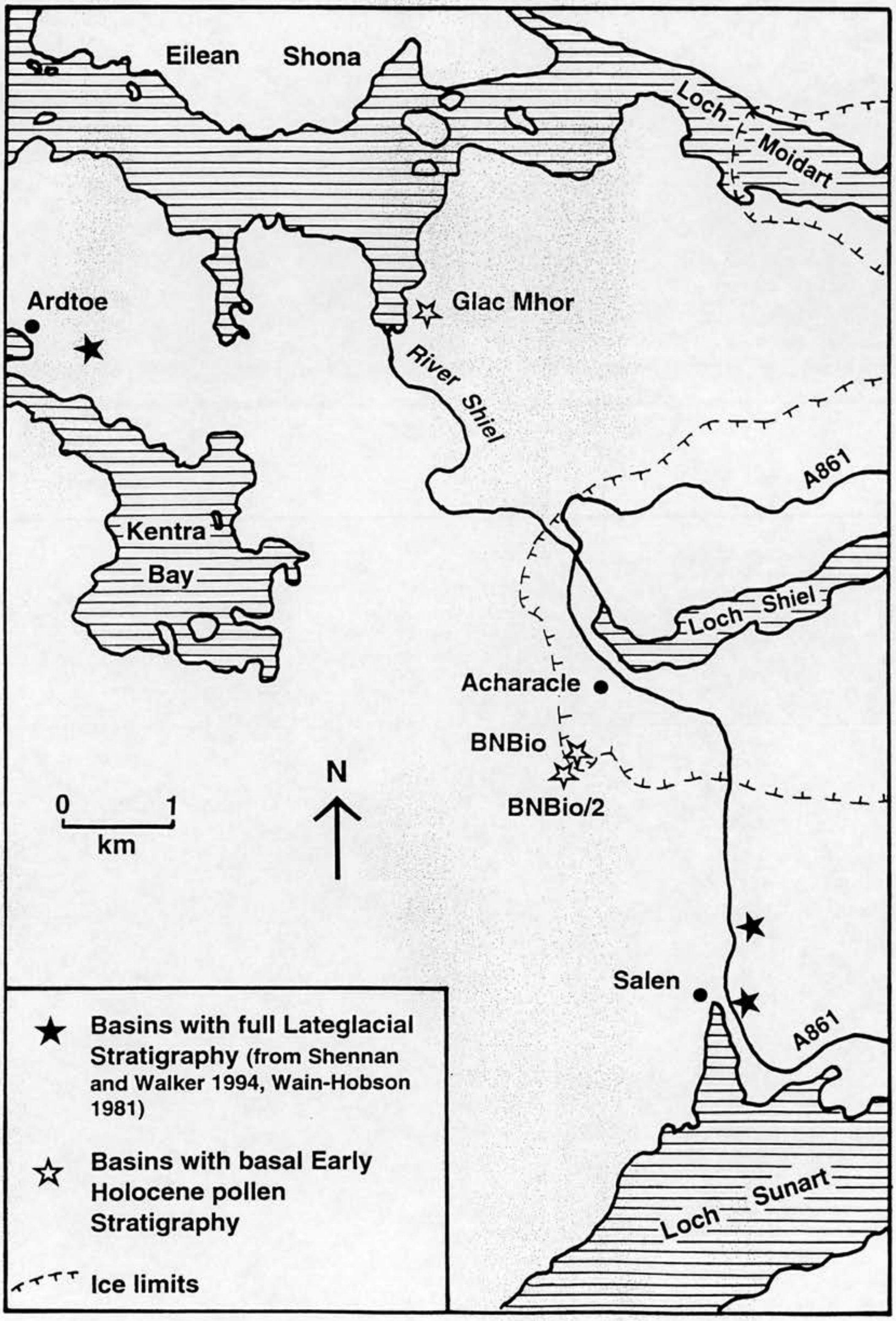


Fig 5.4 Location of depositional basins investigated

In this study several enclosed depositional basins, both inside and outside the inferred ice limits were investigated and the basal sediments examined, in order to provide dating controls on the ice limits mapped here. Cores of basal organic sediments were recovered from three basins in the lower Shiel trough, one inside the inferred limit and two outside. The locations and the core stratigraphies are shown in Fig 5.4 and Table 5.1.

Table 5.1 Core stratigraphies.

BNBIO		BNBIO/2		Glac Mhor	
Depth - cm	Sediments	Depth - cm	Sediments	Depth - cm	Sediments
0 - 570	Peat	0 - 235	peat	0 - 715	peat
570 - 591	Gravels and sands	235 - 261	fine herbaceous peat	690 - 714	fine, pale brown fibrous peat, grading to....
		261 - 279	gyttja	714 - 723	amorphous peat / fine detrital mud with mica
		279 - 335	laminated grey silts, sands and clays, occasional angular clasts	723 - 733	greeny grey micaceous silty clay with plant fragments, grading to....
		335 - 378	grey, coarse silt/ fine sand	733 - 743	greeny brown fine organic detrital mud with mica
				743 - 749	slightly paler, fine clayey detrital mud
				749 - 766	brown fine detrital mud, highly organic, with rare mica
				766 - 780	grey structureless clay/ silty clay
				780 - 815	grey clay
				815 - 820	silty and sandy grey clay

BNBIO is a bedrock basin of 60,000m^2 near Bealach nam Biodag on the rockbar across the mouth of the Shiel trough (NM 676 672). It lies ~0.5km north, and inside the linear moraine that is interpreted as marking the maximum of the Shiel glacier. The bedrock basin is 594cm deep, and contains basal silty clays over gravels with no tripartite litho-stratigraphy (Table 5.1).

BNBIO/2 (NM 677 668) is a smaller 6,000m^2 linear basin immediately adjacent to, and outside the moraine marking the ice maximum. It is enclosed by this moraine on the NE side, and by a bedrock slope to the SW. The corer penetrated to 378cm, but bedrock was not reached. The lower 100 cm of this core consisted of massive and laminated grey silty clays, overlain by gyttja and peat (Table 5.1).

Glac Mhor (NM 664 711) is a basin of 12,500m^2 between the mouths of the Moidart and Shiel troughs and is enclosed by bedrock slopes. It is outside the inferred limits of the Shiel and Moidart glaciers, an assertion corroborated by the presence of a high raised beach at 20-

30m O.D. east of the basin (see Section 5.3.3). A maximum depth of 820m was recorded, and the basal sediments retrieved were a silty sandy grey clay overlain by organic detrital mud.

Other basins outside the reconstructed LLS limits showed no litho-stratigraphic evidence of the tripartite Lateglacial sequence, and two inside the ice limit showed no evidence of early Holocene sedimentation. These basins do not provide evidence relevant to this study, and their locations are listed in Appendix 5.

At both BNBIO and BNBIO/2 the pollen stratigraphy of the basal organic sediments shows that they almost certainly date from the earliest postglacial (Tipping 1994, in Appendix 5), and thus have not been glaciated since the LLS. At site BNBIO/2, immediately outside the Shiel terminal moraine, this early postglacial sequence of gyttja and peat overlies 100cm of minerogenic clays. This unusually thick layer of laminated clays is consistent with the interpretation that the ice front was in close proximity, supplying large volumes of sediment to the basin, and the fact that they lie beneath earliest postglacial sediments suggests that this ice limit occurred during the LLS. At site BNBIO ~300m inside the moraine, early postglacial peats overlie a thin sequence of basal sands and gravels, a sequence typical of sites that lie within LLS ice limits. As sediment accumulation at this site also commenced in the earliest postglacial, the temporal coincidence with the decay of LLS glaciers again suggests that accumulation began as a result of LLS ice retreat, although the absence of Lateglacial sediments cannot prove that this was the case (Tipping 1989).
A tripartite litho-stratigraphy at Glac Mhor, which lies 3km outside the limit of the reconstructed Shiel glacier, proved to contain only an extended postglacial pollen sequence (Mc Culloch pers. comm., sequence shown in Appendix 5), with no evidence for the Interstadial vegetation succession. The clay layer at ~720 - 730cm in the core may be due to a mass movement event inwashing minerogenic sediments from the slopes around the basin at some time in the Holocene.

An LLS age for the ice limit reconstructed here is thus supported by the presence of earliest postglacial basal organic sediments at both BNBIO and BNBIO\2, which suggests that these sites were affected by LLS ice. The presence of Interstadial sediments outside the reconstructed limits confirms that LLS ice did not extend to these sites.

5.3.2 Radiocarbon dating

There are five relevant radiocarbon dates from within this area. Three are from basal postglacial peats in the basin investigated by Wain-Hobson (1981) at Salen, and gave dates of

9,796±75 yrs BP (Wain-Hobson 1981), 10,093± yrs BP and 9,483±70 yrs BP (cited in Wain-Hobson 1981). These give a minimum date for the LLS minerogenic inwash that preceded the accumulation of organic deposits at this site. The basal interstadial/ stadial boundary at this site was dated to 10,643±75 yrs B.P., but this sample may have been contaminated by younger carbon (Wain-Hobson 1981). A date of 8,320±20yrs B.P. was obtained from the basal organic sediments of a raised bog on top of the proglacial outwash fan at Kentra (Shennan et al. 1994b), showing that the fan was deposited before the early Holocene.
An alternative means of dating the ice cap is available if a clear ice limit can be traced from Western Lochaber to somewhere where the limit is dated. The ice limit shown here joins with that mapped by Thorp (1986) in the area immediately to the east, and he traced this limit south to the mouth of Loch Creran. Here, the mean of several radiocarbon dates of marine shells incorporated into morainic deposits is 10,025± 65 yrs BP (Peacock et al. 1987), indicating that the glacial maximum marked by this limit occurred after this time.
These two sets of radiocarbon dates in conjunction suggest that the ice cap developed during the LLS.

5.3.3 Raised marine evidence

Comparison of the locations of raised marine features at different altitudes, with the radiocarbon dated sea-level curve established by Shennan (1994) for the north west of this study area also agrees with the LLS ice cap reconstruction. Shennan's dated curve shows that MSL has not been above ~7m O.D. since the Lateglacial Interstadial in north west Western Lochaber (Fig 1.8). The age of the reconstructed ice cap may thus be estimated by examining marine limits inside and outside the ice maxima in Lochs Shiel and Ailort, where sea level evidence was **not** used to determine the ice limits. If the ice cap dates from the LLS, the marine limit should be at or above 7-8 m O.D. outside the ice maxima, and the marine limit inside ice maxima should be at or below 7-8m O.D. Fig 2.21 and Table A.3.1 in Appendix 3 show that the marine limit inside the ice maxima in these lochs is 7.7m OD and the limit outside is 30 - 40m OD, which agrees with an LLS age for the ice cap.
The sea-level curves suggest that the limit around Loch Shiel occurred during the LLS. The evidence provided by the lowest measured altitudes of subaerial fluvioglacial outwash sediments outside the ice limit at the mouth of Loch Shiel suggests that when the gravels were deposited contemporary MSL was <4.5m O.D. Comparison with Shennan's curve shows that the gravels must therefore have been deposited either in the latter half of the LLS, or in the late Holocene. As the surface of these gravels has been affected by subsequent marine action which can only have occurred during the Postglacial Marine Transgression (Shennan et al. 1994b), it follows that the second alternative age cannot be correct. The Shiel glacier was

therefore in a maximum position depositing proglacial outwash gravels during the latter half of the LLS.

The coastal proglacial outwash fans which show subaerial surface features, such as kettleholes and channels, provide additional chronological constraints. The maximum surface altitudes of these fans are all between 12.5 and 27m O.D. This is above palaeosea-level altitudes during the LLS, yet below those maintained during the Late Devensian (Section 1.4, Table 2.3). This suggests that the fans were formed during the LLS.

All these three independent lines of chronological evidence in all parts of Western Lochaber thus agree that the reconstructed ice cap existed during the LLS.

5.4 The LLS deglaciation pattern

There is clear evidence for the pattern of glacial retreat in parts of Western Lochaber, and this evidence can be used to suggest the processes involved.

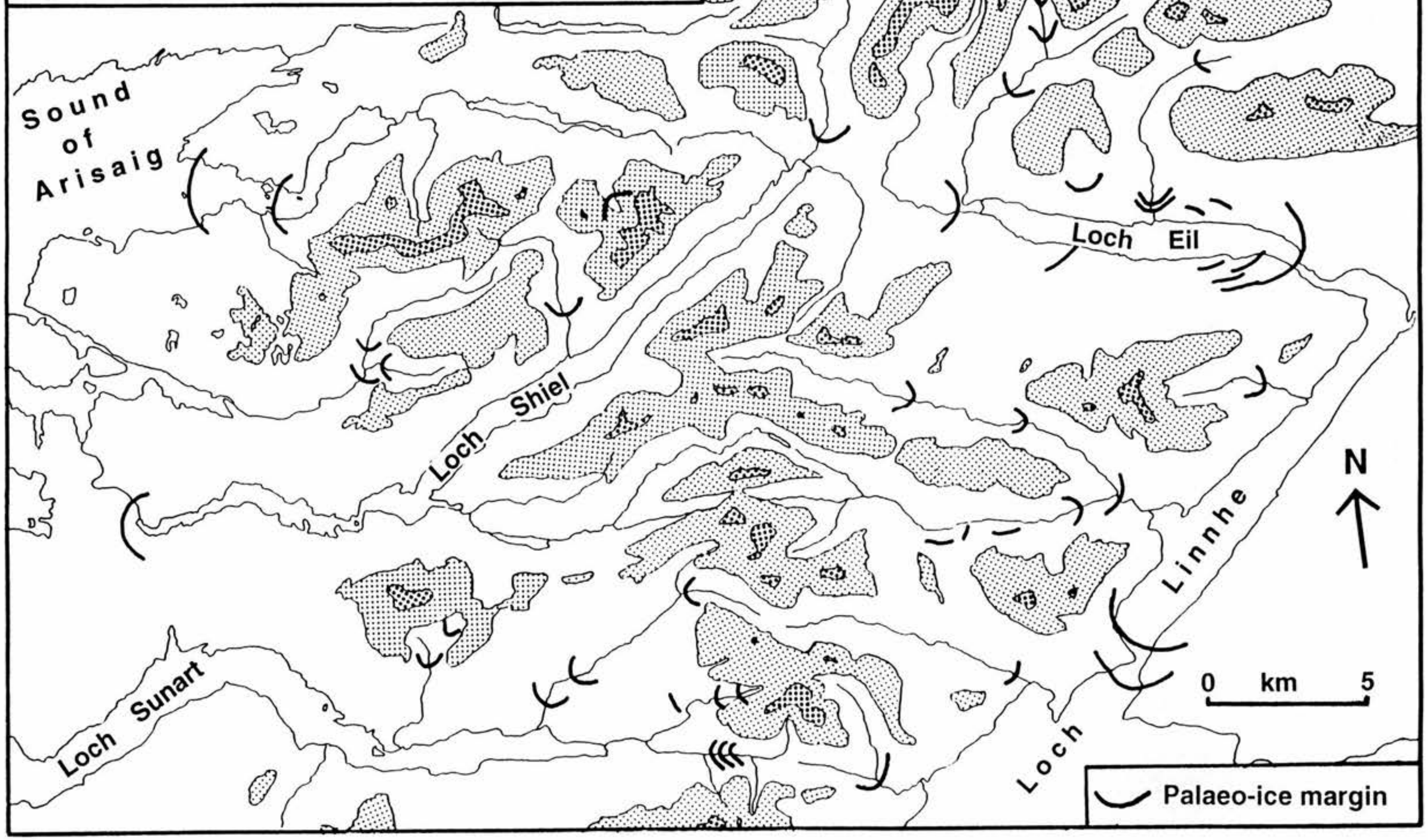
The deglaciation pattern shown in Fig 5.5 has been reconstructed by mapping depositional evidence for terminal and lateral ice margins at retreat stages, inside the maximum ice limits. The best evidence for retreat stages in Western Lochaber is in the form of the fluvioglacial outwash fans described in Section 2.4. It is likely that fans with clear ice contact slopes and those which contain large volumes of debris record former ice marginal positions. These are found around the former sea loch troughs, in the locations shown in Fig 2.1.

At almost every location where there is a narrowing or a shallowing in a sea loch there are glacial outwash deposits associated with it. Outwash deposits are also found at the heads of many of the lochs and where they are joined by side glens. The terminal positions indicated by these deposits all suggest that ice retreated up the sea lochs to the glens and mountains at their heads.

There is also evidence for retreat stages on land. Both large moraines, for example those at the head of Loch Eil and the mouth of Glen Suileag on Loch Eil, and smaller linear moraine ridges found amidst 'hummocky moraines' are likely to give evidence for former ice front positions, as argued by Bennett and Boulton (1993a,b) and Benn (1992). Examples of these are common around Loch Eil, and in western Glen Tarbert. These moraine fragments all show that the glaciers retreated up glens to the mountains at their heads (e.g. Greene et al. 1994).

In some instances drift terraces, ice contact slopes and meltwater channels incised into drift record former ice front positions. These features are widespread around Loch Eil, and in Glens Fionnlaighe, Suileag and Scaddle, and all indicate ice thinning and retreating up valleys.

Fig 5.5 Deglaciation of Western Lochaber



In addition to retreat patterns, tentative inferences about relative rates of retreat in different locations may be made. The size of proglacial and ice marginal deposits is dependent on the rate of sediment supply to the former terminus, which is logarithmically related to catchment size (Powell 1991), and the length of the glacial stillstand at that location. Whilst these parameters cannot be directly determined, the large size of many outwash fans, compared to the volumes of other glacial deposits in Western Lochaber, suggest that they mark the sites of stillstands during retreat. The largest fans are found at Corran in the Linnhe trough and in the lower Shiel trough. These had the largest catchments according to the reconstruction shown in Fig 5.1, yet the volume of outwash deposits is many times greater than in other troughs. Further support for a glacial stillstand at Corran is provided by the seismic stratigraphic records. Firstly, the arrangement of the LLS glacimarine unit suggested a debris source around the Corran narrows. Secondly, there is a moraine bank ~27m high just south of the narrows. The size of this feature suggests that it marks the location either of the ice maximum or of a stillstand position. As these positions are topographic pinning points (see Section 6.2.2), it is also possible that these sites have been the sites of previous glacial stillstands. If so, the large size of some fans may be partly due to LLS overriding or reworking of pre-LLS glacial stillstand deposits. This possibility may be supported by the degree of clast roundness in the outwash deposits around lower Loch Shiel (Section 2.3.3).

On land, large moraine ridges are likely to mark a glacial stillstand or readvance, and the size of the deposits may be crudely indicative of the length of stillstand or readvance, if it is assumed that sediment supply to the terminus was fairly constant spatially and temporally. It is possible, for example, that smaller ridges represent annual winter readvances (c.f. Bennett 1991, Boulton 1986).

If this reasoning is correct, the large size of the moraines around Loch Eil suggests they may mark stillstand terminal positions during retreat related to withdrawal onshore. In several glens, stillstands may have occurred at rock bars and at confluences of troughs, as large moraine ridges and accumulations of moraine are commonly found in these locations (e.g. Gleann Feith 'n Amean and Glen Scaddle, respectively).

Most of the clear moraine lineations are found in the middle and upper parts of the glens, and are absent from the lower sections. In Glen Scaddle, for example, there are only sporadic glacial deposits until 4-6km from the mouth. Upvalley from this point drift deposits are thick and moraine fragments abundant. This pattern may suggest that retreat proceeded rapidly through the lower glens and more slowly in the upper glens (c.f. Benn et al. 1992). Similarly, the poor deglacial evidence in the southern and western glens due to the lack of drift and moraines may reflect rapid deglaciation in these glens. However, variations in the spatial and temporal supply of debris to the ice termini cannot be ruled out as other factors helping to explain these observations.

The evidence for deglaciation thus suggests that ice retreated up the glens towards the ice sheds and the mountains at their heads. Stillstands in the lochs occurred around shallows, narrows and at the mouths of glens. On land smaller scale stillstands may have occurred at glen confluences and rock bars. Deglaciation of the upper glens may have been slower than that of the lower glens, or debris supply rates may have been enhanced during the final stages of deglaciation.

The retreat pattern inferred from moraines, outwash fans and areas of aligned and chaotically arranged hummocky moraines contrasts with that mapped by Bennett (1991) and Bennett and Boulton (1993a,b). Comparison both of the distribution of moraines shown in Fig 2.1 and of the deglaciation pattern shown in Fig 5.5 with the retreat stage map constructed by Bennett (1991) on the basis of air photograph evidence for hummocky moraine lineations in the same area (Fig 1.8) shows that the maps correspond poorly. In 14 of 25 glens in Western Lochaber, Bennett's depiction of the presence or alignments of hummocky moraine differs from those found in this study. Bennett found numerous aligned moraine ridges and fragments, but no examples of chaotically arranged moraines. In some glens this study has found clear evidence of aligned moraines where Bennett maps none (Glens Moidart) or few (e.g.Cona glen). In other glens Bennett maps numerous linear moraine ridges and fragments where this study found none or very few (Glen Scaddle, around Loch Eilt, Glen Hurich and Coire an Lubhair). In Glens Finnan, Dubh Lighe and Fionnlaighe Bennett maps numerous moraines aligned in a manner suggesting retreat towards the mouth of these glens, whilst this study found few moraines, but these were oriented to suggest retreat towards the heads of these glens (Greene et al. 1994). In eastern glen Tarbert detailed mapping undertaken in this study shows moraines which are chaotically arranged, whereas Bennett marks a suite of aligned ridge fragments. Finally, in the glen north of Strontian glen there are a series of glacial flutes which cross the glen diagonally so that they trend uphill on the southern flank of the glen; Bennett records a suite of ice marginal positions along these ridges. It is suggested here that these discrepancies may arise from the contrasts in methodology. Bennett relied on air photograph interpretation while this study also emphasised field checking. The retreat pattern produced by Bennett and Boulton (1993a,b) also differs from that produced here as they rarely used outwash fans as evidence for palaeo-ice marginal positions, whereas these are emphasised in this study.

5.5 Summary

The positions of moraines, trimlines and seismic evidence all support the ice limits presented here. The reconstructed ice cap shows a thick ice plateau centred on the mountains in the centre and north of the study area, with outlet glaciers flowing down the loch troughs to the sea. All existing evidence for the age of this ice cap is consistent with the interpretation that it existed during the LLS. During deglaciation the LLS glaciers retreated back up the glens, towards the mountains. Retreat up the sea lochs may have been punctuated by stillstands at narrows and shallows in the lochs, and at the mouths of glens. Other ice marginal positions during retreat suggest smaller stillstands may have occurred in the upper glens, at valley confluences and at rock bars.

Chapter 6 - Discussion

6.1 Aim

The aim of this Chapter is to explore the wider implications of the reconstructed LLS ice cap. These include inferences about controls on ice cap dynamics based on the nature of glacial evidence in Western Lochaber, reconstructed Equilibrium Line Altitudes, and the nature of Holocene geomorphic activity.

Part I

6.2 Controls on glacial erosion, deposition and ice dynamics

6.2.1 Introduction

The spatial pattern of glacial erosional and depositional features is distinctive at two different scales (Table 6.1);

A. There is a clear contrast between the types of glacial evidence present in the south and west of Western Lochaber, compared to that in north east Western Lochaber and Eastern Lochaber.

In the former area there is an abundance of signs of glacial scouring and few glacial deposits, with a notable absence of terminal and lateral moraines relating to the glacial maximum in the sea loch troughs. Of the glacial deposits present, large volumes occur in fluvioglacial outwash fans around the sea lochs. By contrast, in Eastern Lochaber there are widespread glacial deposits including large terminal and lateral moraines relating to the ice maximum, and few signs of glacial erosion (Thorp 1984, Sissons 1979a).

B. Within troughs in Western Lochaber, glacial deposits are concentrated in the bases of the glens, and in hollows or basins on the glen sides. Till is common in tributary glens at right angles to the strike of the main glens. Glacial till and hummocky moraines are most prevalent in the upper halves of the glens. Fluvioglacial deposits are locally very thick, and are found at narrows and shallows in the lochs, and at the heads of lochs or where tributary glens join sea lochs. Glacimarine LLS deposits are thickest in the proglacial maximum submarine basin in Loch Linnhe, with only thin, sporadic deposits in the basin through which the glacier retreated. Glacial scouring is common around spurs and rock bars in glens, the sections of glens which are narrow and steep sided, and in the lower sections of glens which join sea lochs.

Table 6.1 Spatial contrasts in the distribution of glacial evidence in Lochaber.

Scale	Erosional features	Till and moraine deposits	Proglacial fluvioglacial deposits	Glacimarine deposits
Between Eastern and Western Lochaber	south and west of Western Lochaber	north east of Western Lochaber and Eastern Lochaber	around sea lochs and in east of Eastern Lochaber	sea lochs
Within troughs	spurs, narrow steep sided sections, along main mountain axes	mid and upper glens, base of glens, hollows, cols and side glens	shallows and narrows, heads of lochs, mouths of tributary glens	basin outside the former ice maximum (within fjords)

These field relationships reflect controls on patterns of glacial erosion and deposition at the two different scales. Table 6.2 shows the factors which are likely to have been important controls on glacial erosion, deposition and dynamics in Lochaber.

Table 6.2 Controls on glacial dynamics in Lochaber

	W-E Lochaber contrasts	Within trough contrasts
Tidewater glacier controls	✓✓	✓✓
Topographic controls	✓	✓✓
Geological controls	? ×	✓
Legacy of previous glacial episodes	✓✓	✓

6.2.2 Tidewater glacier controls

Tidewater glaciers differ in behaviour from land terminating glaciers due to the importance of iceberg calving at their termini as the dominant means of ablation. Calving velocities increase with increasing water depth (Brown et al. 1982), and the efficiency of calving as a means of mass loss results in the distinctive dynamical characteristics of glaciers which terminate in deep water (>~80m).

Iceberg calving results in extending, fast flow in the lower reaches of glaciers (Meier and Post 1987), particularly if the glacier terminus is in deep water and/or in a wide trough (Powell 1991). Rapid calving at the glacier snout means that an ice 'draw down' effect may enhance velocities upglacier, possibly aided by the presence of soft, easily deformable glacimarine sediments on the sea bed (Powell 1991, Boulton 1990). Due to the high iceberg calving rates, advance through deep water can only occur slowly, at a rate dependant on the rate of sediment supply to the terminus and water depth (Powell 1991), as the ice front advances on a submarine moraine shoal which lowers the effective water depth (Post 1975, Mayo 1988). If the glacier terminus moves back from the moraine shoal into deep water, retreat rates are rapid. Tidewater glaciers only attain stable stillstand positions during retreat at 'pinning points' where the iceberg calving rate is reduced. Calving rates decrease when a glacier retreats to a location where the trough is shallower and/or narrower than elsewhere, or back onto land (Mercer 1961). Rapid retreat up sea lochs may result in temporary stabilisation of the ice fronts when they retreat back on land and are left with a positive mass balance for a while (c.f. Mann 1986, Mayo 1988). These calving glacier controls apply to glaciers with both tidewater and freshwater termini, but the effects are more pronounced in the case of the former (Warren 1992).

Tidewater glacier dynamics result in patterns of erosion and deposition which are different to those of land terminating glaciers. These are illustrated schematically in Fig 6.1. If it is assumed that ice flux is an important control on rates of erosion (Andrews 1972), erosion may be high where ice velocities are high (Shoemaker 1986, Boulton 1990). In a terrestrially based glacier, velocities rise to a maximum at the ELA, and decline towards the snout, hence erosion may be most effective in the middle reaches of these glaciers. Deposition may be concentrated below the ELA, rising to a maximum at the terminus, as a result of ice flow lines and compressive flow in the lower reaches (Fig 6.1, glacier A).

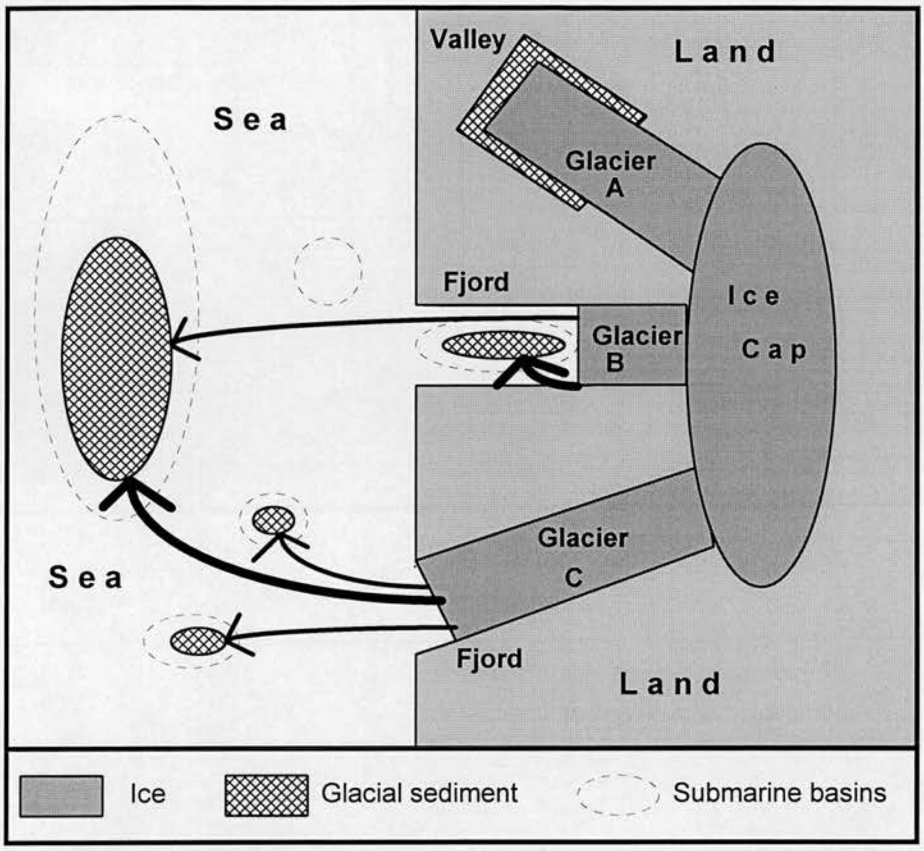


Fig 6.1 Relationship between location of ice terminus and location of glacigenic sediment.

Glacier A has a terrestrial snout and glacial sediment is deposited at the ice margins below the ELA.

Glacier B has a tidewater terminus within a fjord, and glacial sediment is concentrated in the proglacial submarine basin.

Glacier C terminates at the mouth of a fjord and glacial sediment is located in distal submarine basins offshore, and in any deep basins closer to land.

In contrast, calving glaciers have high ice velocities in the lower reaches, so that debris may be efficiently evacuated to the glacier snout. In addition, extending flow in the lower reaches means that upwards movement of debris-rich basal ice is unlikely to occur, resulting in smaller supplies of ice marginal debris than on land based glaciers (Dowdeswell 1986). Deposition in this situation is concentrated at the ice terminus below sea level, mainly in the form of material discharged by meltwater streams (Powell 1984). These glacimarine deposits are mostly confined to the basin immediately seawards of the ice terminus (Powell 1981, 1983, Elverhoi 1984, Syvitski 1989)(Fig 6.1, glacier B). If the glacier terminates at a fjord mouth where there is no proglacial basin glacimarine sediment may be transported by icebergs and ocean currents to the continental shelf and slope (Powell 1991) (Fig 6.1, glacier

C). As tidewater glaciers must advance through deep water on moraine shoals, where large submarine moraines are preserved, these are likely to mark ice maxima or readvances.

Rapid retreat means that only thin spreads of glacimarine deposits are found in deep retreat basins, a pattern exacerbated by the contracting size of the glacial drainage basin (Powell 1991). Rapid retreat may be punctuated by stillstands at pinning points, such as shallows, the heads of lochs and the mouths of side glens. Here proglacial outwash fans may build up on the sea bed and aggrade to sea level and above (Powell 1988, 1990). Erosion, in these high velocity, high ice flux glacier systems, is intense and may continue as far downglacier as the ice front (Powell 1991, Boulton 1990).

The evidence in Lochaber conforms with these observations. There are clear contrasts in the distribution of erosional and depositional features between the catchments of former LLS glaciers that flowed south and west advancing through tidewater, and those that flowed east advancing over land. Where glaciers had tidewater termini, there are widespread terrestrial signs of erosion and few glacial deposits around the loch troughs, particularly the deep sections, and close to the LLS maxima (c.f. Fig 6.1, glaciers B and C). In the eastern half of the former LLS ice cap, termini would have been terrestrial for much of the advance period, although at the ice maximum some termini were freshwater (Sissons 1979a). Glacial deposits in Eastern Lochaber are widespread, partly because outwash fans which were deposited in a subaqueous environment in ice dammed freshwater lakes are now visible subaerially, but drift deposits are also more ubiquitous and thicker (c.f. Fig 6.1, glacier A).
Terrestrial glacial till deposits are mostly found in the mid and upper troughs, away from the lochs, and in the former ablation areas of small corrie glaciers. These are situations in which glaciers would have had terrestrial termini.

The evidence discussed in Chapter 4 shows that within troughs, offshore glacial deposits are concentrated in the locations predicted by tidewater glacier theory. Glacial maxima in Lochs Nevis and Ailort are marked by submarine moraine shoals (Boulton et al. 1981), and this may be the case in Loch Linnhe. The Nevis and Ailort glaciers reached the mouths of these sea lochs. It is likely that glacimarine sedimentation from these glaciers was concentrated in basins on the continental shelf, due to strong tidal currents over the relatively shallow waters outside these ice termini (c.f. Fig 6.1, glacier C). Seismic evidence in Loch Linnhe shows large volumes of LLS glacimarine debris in the basin in which the glacier terminated, which is typical where such a basin exists (c.f. Fig 6.1, glacier B). There is little evidence of LLS glacimarine sediments elsewhere in the loch, including the retreat basin, which suggests that retreat through the upper loch may have been rapid. The stillstand retreat positions marked by outwash spreads and moraines around the sea lochs correspond exactly with topographic pinning points; narrows and shallows in all the sea lochs are associated with large outwash fans, suggesting that these are the locations of stillstands during retreat (Greene 1992). Thus, for example, stillstand outwash deposits are found at Corran and Annat on Loch Linnhe, the only two places where there are sills in the Eil / upper Linnhe trough (Fig 6.2). Outwash fans are not associated with lateral moraines, or drift limits, which suggests that most deposition was concentrated proglacially, at the ice front. Temporary ice front stabilisation on land following rapid retreat up sea lochs is suggested by outwash spreads at Aladale on Loch Shiel, at Strontian and Laudale on Loch Sunart, and at Inverscaddle on Loch Linnhe, and moraines at Fassfern and Kinlocheil on Loch Eil (Figs 5.5, 6.2).

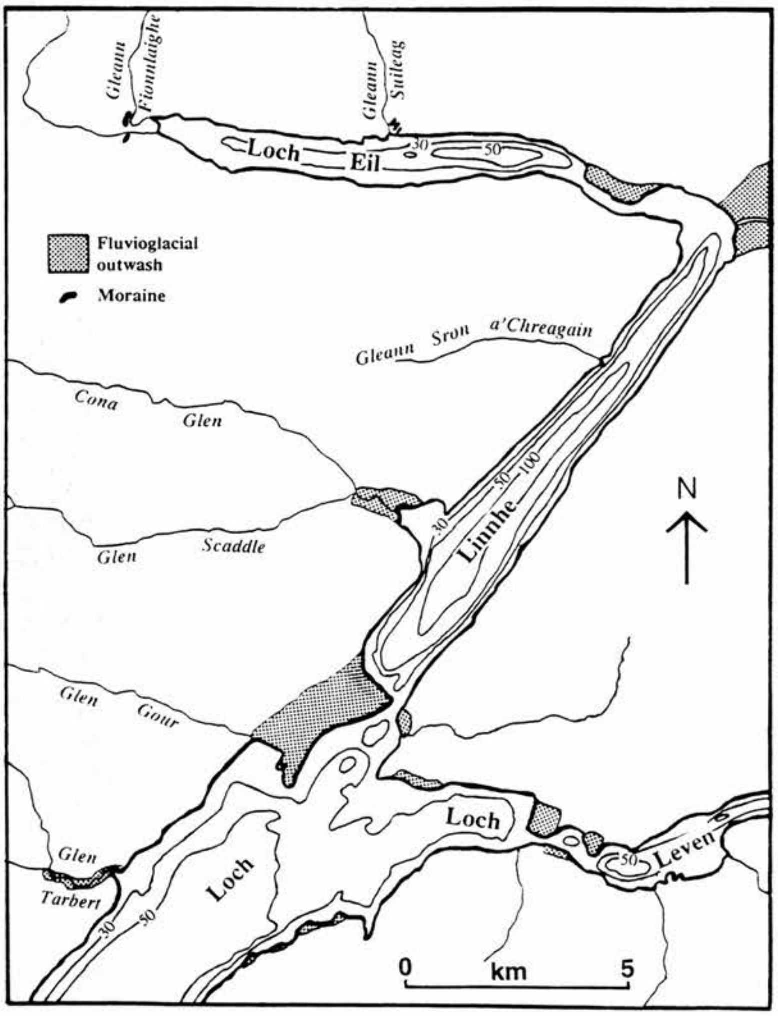


Fig 6.2 Bathymetry and evidence for LLS glacial stillstands in Loch Linnhe.

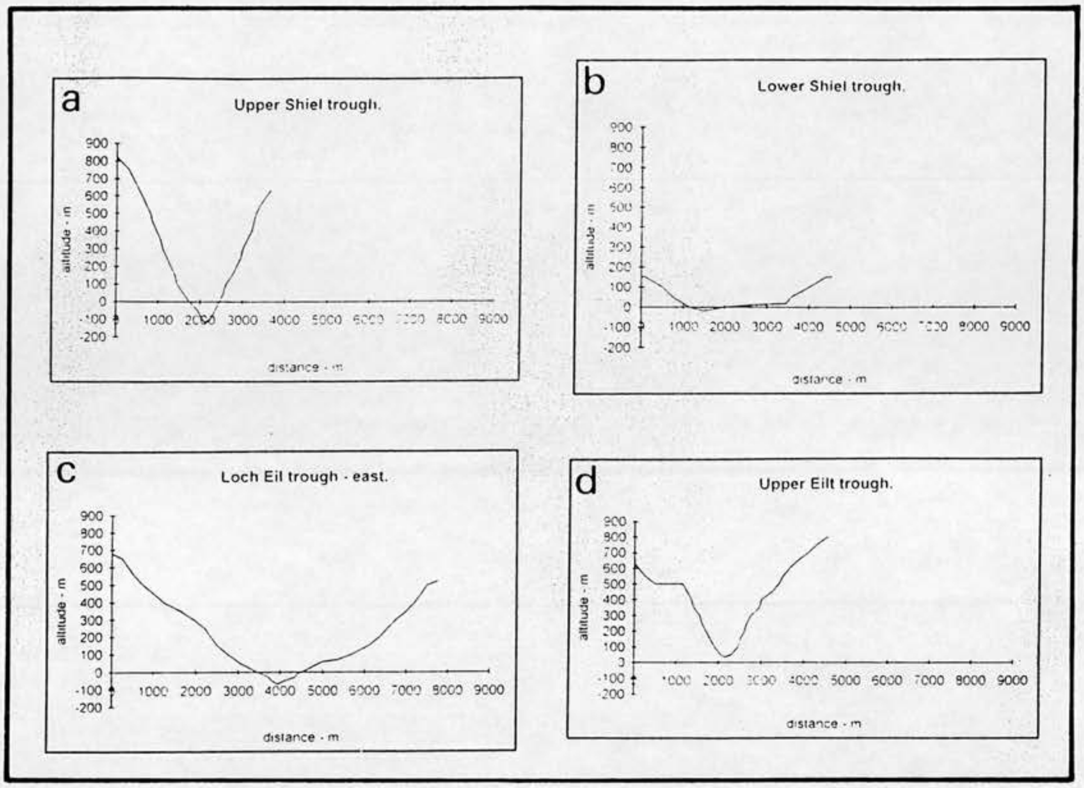


Fig 6.3 Trough cross-sections.

The control of trough geometry on calving rates and thus ice dynamics may also be illustrated by an examination of the location of fluvioglacial deposits around Loch Shiel. The lower half of the trough is a broad, low sided basin, and the loch depths reach a maximum of 30m (Fig 6.3). This part of the trough contains great expanses of fluvioglacial deposits pitted with large kettleholes on either side of the loch. This suggests that retreat in this shallow section of the Loch may have occurred slowly, allowing time for the accumulation of these deposits. By contrast, the upper half of the Shiel trough is narrow, steep sided, and flanked by high mountains (Fig 6.3). The loch reaches depths of over 120m. There are no significant glacial deposits of any form around the loch shores. This suggests that final retreat may have occurred rapidly up this deep section of the loch, not allowing time for debris accumulation at the glacier margins.

The glacial evidence around the shores and on the floors of the former sea lochs thus shows that the deglaciation pattern and ice behaviour were influenced by tidewater glacier dynamics and the location of topographic 'pinning points' in the lochs. The field evidence thus supports the idea that tidewater glacier dynamics have been important controls on LLS ice behaviour, and affected the distribution of glacial erosional and depositional features at both spatial scales.

The importance of iceberg calving in influencing icecap morphology and dynamics is also shown by the results of the modelling experiments discussed in Section 5.2.3. Modelling ice cap growth with and without the effect of calving produced different ice cap morphologies. These results emphasise the importance of the Linnhe glacier as an icecap outlet. If the base of this trough was above sea level so that the glacier did not have a calving terminus, the model suggests that ice would have advanced down the trough as far as the south of Lismore, and that ice flow across Western Lochaber would have been radially away from the Ben Nevis area. By contrast, incorporating the effects of calving into the model produces an icecap of similar morphology to that reconstructed here. Calving may thus have been an important control on ice dimensions and ice flow directions.

6.2.3 Topographic controls

The patterns of erosion and deposition in Lochaber suggest that topography may have been an important control on ice dynamics. Topography plays an important role in influencing ice flow patterns. Firstly, at a regional scale, the altitude and morphology of mountains and their interaction with climate place constraints on the growth, decay and potential size of a glacier (Kerr 1993). Secondly, ice flow may be different in troughs of different geometry. If it is assumed that there is a general relationship between ice flux and debris transport and evacuation, then high velocities may be associated with enhanced rates of erosion (Andrews 1972, Shoemaker 1986). Where glacier flow is confined by narrow, steep sided trough walls, or by rock bars and spurs, velocities are enhanced, and ice fluxes increase. Velocities are also increased where there are convexities in the bed long profile. Conversely, flow in broad, unconfined troughs may involve lower sliding velocities and ice fluxes, reducing the potential for erosion (Shoemaker 1986), and enhancing that for deposition.

At a regional scale, the arrangement of erosional and depositional evidence suggests that topography was a crucial control on the growth and decay of the ice cap. Ice directional indicators and depositional evidence for ice termini show that ice accumulated in the main mountain axes and flowed from the hills down troughs towards the coasts to the south and west. Depositional evidence for retreat stages shows that ice decayed back from the coasts to the mountains at the heads of glens.

Topography has also played an important role in determining patterns of erosion and deposition at smaller spatial scales. Narrow, steep sided troughs in Lochaber, such as the upper Shiel and Eilt troughs, show numerous signs of glacial erosion. Evidence suggests this was particularly intense in certain locations, for example where there are rock bars and spurs projecting into troughs (e.g. Glens Gour and Strontian), and around a bedrock step in the longitudinal profile of the Eilt trough. In contrast, shallow and broad troughs, such as the lower Shiel trough, the Eil trough and many troughs in Eastern Lochaber, have few signs of intense erosion (Fig 6.3). In these locations slopes are often mantled with glacial deposits. Within troughs, thick glacial till occurs in side glens (e.g. Glen Tarbert, Cona Glen), in the lee of obstructions to the main local ice flow directions or in hollows on the sides of glens (e.g. Glen Hurich). This may be a further result of topographic influences on ice velocities; in locations such as side glens and the lee of spurs, velocities may be lower than elsewhere, and debris may not be evacuated as efficiently as in the main troughs.

Topography may also have influenced the locations of moraines and accumulations of glacial deposits in troughs. Numerous terminal and retreat moraines in Western Lochaber are at least partly superimposed upon bedrock ridges (e.g. in Glen Feith n' Amean, south of Beinn Resipol, Coire na Cnamha, the terminus of the Shiel glacier). Moraines are also often concentrated at the confluences of troughs (e.g. Glen Scaddle). It is thus likely that glacier terminal positions were influenced by local topography creating stable 'pinning points' (c.f. Mercer 1961, Andrews 1975).

6.2.4 Geological controls

Geological controls on the glacial erodibility of bedrock are often important influences on the distribution of glacial deposits (e.g. Bennett 1991, Boulton et al. 1991). Where bedrock is easily erodable, glacial deposits may be more prevalent down-glacier. Glacial erodibility depends on the jointing pattern and rock discontinuities (Le Coeur 1994, unpublished). Interestingly, there appears to be no correlation between the main geological units and the location of glacial deposits in Western Lochaber. Over most of the area bedrock may be fairly uniform in terms of glacial erodibility, although there have been few detailed studies of bedrock jointing patterns. The fissile micaceous semipelitic and pelitic schists are well jointed, and the coarse grained acid intrusions may be more easily eroded than schists and gneisses, yet no thickenings of till were observed at, or downglacier of, outcrops of these types. Stratigraphic subdivisions of the rocks of the Loch Eil division of Moine schists are presented by Stoker (1985) and Strachan (1985). Comparison of these results and the glacial geomorphology shown in Fig 2.1 suggests that scoured slabs are most widespread, and superficial drift deposits least prevalent, on the massive, coarse grained quartzites, siliceous psammites and gneisses in this area. Scoured slabs are less widespread where bands and units of micaceous semipelites and pelites are present. Detailed geological work elsewhere in Western Lochaber is needed to see if this correlation holds over a wider area, and to determine whether this is a function of lithological resistance to glacial erosion and /or to postglacial weathering.

6.2.5 Glacial history of Western Lochaber - Legacy of Devensian dynamics

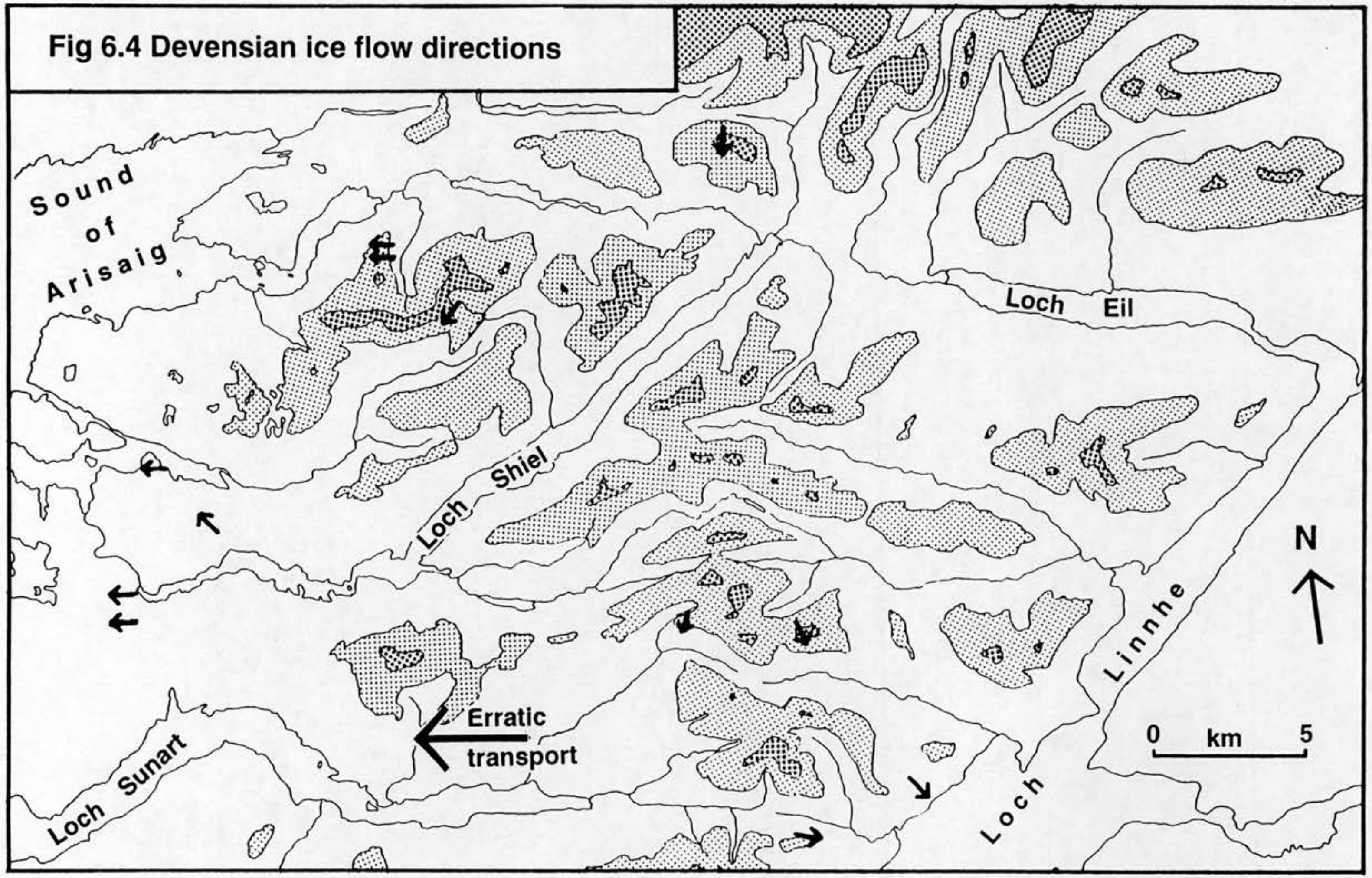
Several lines of evidence suggest that the distribution of glacial erosional and depositional features in Lochaber is, at least in part, a relict pattern inherited from the glacial record left during the Devensian, and maybe previous, glaciations.

Spatial distribution of glacial erosional and depositional features.

Evidence outside and above the LLS ice limits shows an identical east west contrast in the distributions of erosional and depositional features to that observed for LLS glacial features. There are numerous smoothed slabs both outside the LLS limits in the west of the area, and above the limits, especially around the main mountain axes in Ardgour and Moidart, and in steep sided, narrow troughs. This suggests that the ice sheet was warm-based and that glacial erosion predominated throughout Western Lochaber during the Devensian glaciation. The sediment thickness transect showed that there are large amounts of Late Devensian glacial deposits outside the eastern margin of the LLS ice cap. These are in the form of drift, moraines, and large valley side outwash terraces. There are very few terrestrial glacial deposits outside and above the LLS limits in Western Lochaber.

Although there are few terrestrial Late Devensian deposits, offshore evidence shows thick depositional sequences. Late Devensian ice advance is represented by a clear erosion surface, and local thin pockets of till on the Hebridean continental shelf. Late Devensian deglacial glacimarine deposits are up to 130m thick, and widespread on the outer and inner continental shelves outside the LLS ice limits (Chapter 4, Boulton et al. 1981, Davies et al. 1984, Fyffe et al. 1993, Stoker et al. 1993). Evidence in Loch Linnhe suggests that thick Late Devensian glacimarine sequences in the Kentallen and Shuna basins may have been deposited during a deglacial stillstand around the Corran narrows (Section 4.7.2). Since there is no evidence for terrestrial Late Devensian deposits outside or above the LLS ice limits around Loch Linnhe, or elsewhere in the coastal zone of Western Lochaber, sedimentation during Devensian deglaciation was concentrated at the tidewater ice fronts. These observations suggest that during advance and at the ice maximum the Late Devensian ice sheet had an erosional regime in the west of Scotland. During deglaciation thick depositional sequences were only deposited offshore in Western Scotland. In contrast, terrestrial deglacial deposits are widespread in central and eastern Scotland. It is possible that this pattern reflects the influence of high calving velocities at the western ice sheet margins, in a manner analogous to that which may

Fig 6.4 Devensian ice flow directions



have applied to the LLS ice cap. This lack of Devensian terrestrial deposits in Western Lochaber may mean that there were few glacial deposits to be entrained and redeposited by the LLS glaciers in the west, an additional reason for the lack of LLS glacial deposits in this part of Scotland.

Contrasting ice flow patterns.

There is also clear evidence for Devensian ice flow directions, summarised in Fig 6.4. The striations, roche moutonnées, dispersal of erratics and meltwater channels that lie outside and above the LLS limits support the previously published pattern of westwards movement of ice in Moidart and Sunart, and south eastwards flow in Eastern Ardgour (Bailey and Maufe 1960, BGS 1:50,000 drift sheets 52 and 61, Peacock 1970, Thorp 1987). In Moidart, for example, there are several large meltwater channels trending east-west which cross rock bars across the western glens. In Ardgour, there is high level meltwater channel evidence to suggest that there was some north-south movement of ice across the main east-west ridges, also supported by striations mapped by the BGS (Bailey and Maufe 1960). In Eastern Ardgour, eastwards flow towards the Linnhe trough is supported by striations approximately 3 km from coast in Glen Tarbert. This may be related to rapid ice flow down the Linnhe trough.

Ice flow directions were thus different during different glacial episodes. Ice flow during the LLS was mostly topographically constrained, so that flow directions were usually concordant with the local topography. However, under thicker ice during the Devensian glaciation, ice flow was more independent of local topography, with ice movement over ridges and summits. This contrast is exemplified around Seann Chruach (T1) where Devensian roche moutonnées above the former LLS ice surface indicate westwards ice flow (Fig 2.1), whereas fluted moraine (Figs 2.1, 2.11) and trimline evidence shows that LLS ice flow in the trough immediately east was in a northerly direction.

In Ardgour, ice flow directions were eastwards during the LLS, but south eastwards in Eastern Ardgour, and south westwards in Western Ardgour during the Late Devensian maximum. It is therefore possible that different sized ice sheets in Scotland have been associated with different ice flow directions in this area. This idea is supported by the numerical ice sheet modelling experiments discussed in Section 5.2.3. Under many climatic scenarios and stages of ice sheet growth, the model emphasises the importance of the Ben Nevis area immediately north east of Western Lochaber as a centre for radial ice dispersal, so that ice flow is westwards across Ardgour. By contrast, under some scenarios, for example using imposed temperature depressions of -8^{o}C or -9^{o}C and short model run times, ice flows eastwards across Ardgour, away from smaller centres of accumulation in the mountains of

Western Lochaber. This contrast in ice flow directions in certain areas has interesting implications. The contrast may explain the larger volume of glacial deposits in some of the eastern Ardgour glens, compared to the western glens. Deposits may be trundled up and down these latter glens in different glaciations, whereas they may be more efficiently evacuated in the west by uniform ice flow directions at all times. Contrasting ice flow directions may also explain the distribution of roche moutonnées in Western Lochaber. It is in the western glens, where the Devensian and the Loch Lomond Stadial ice flow directions were the same, that roche moutonnées are present; in Eastern Ardgour bedrock mounds are often smoothed in both directions. This suggests that the Loch Lomond Stadial ice was not of sufficient extent and/or duration to produce roche moutonnée bedrock forms from scratch. Instead, the roche moutonnées in Western Lochaber may be reworked features that were originally formed during the Devensian, or earlier, glaciations. However, these hypotheses remain speculative as the possibility that the roche moutonnée forms are structurally controlled has not been fully investigated (c.f. Gordon 1981).

The field evidence therefore suggests that the Devensian ice sheet was warm based and erosive in Western Lochaber. The fact that even the highest summits show considerable evidence of ice moulding suggests that ice was relatively thick in this area. Main ice flow occurred towards the western calving margin, but the Linnhe trough was also a major ice sheet outlet.

Implications

The persistence of striking west-east contrasts in the distribution of erosional and depositional features outside and above the limits of the LLS ice cap suggests that landscape modification during the LLS may have been limited; the main patterns of scoured and drift-covered landscapes may have been established during previous glaciations.

There are several lines of evidence which support this idea. Firstly, if the interpretation given above for the lack of roche moutonnée forms in Ardgour is correct, this suggests that ice action during the LLS alone was not sufficiently thick or long-lived to produce these forms. Secondly, an examination of the slopes above and below LLS glacial trimlines does not reveal any changes in slope form or angle, suggesting that the glaciers eroded little.

Thirdly, evidence suggests that the 'Lateglacial rock platform' may have been formed prior to the LLS (Section 1.4), yet the feature was not destroyed by the action of LLS ice. This evidence is augmented by observations in Western Lochaber. Most of the locations where the rock platform is present (Fig 2.1) are well within the limits of the reconstructed LLS ice cap (Fig 5.1) and evidence at three locations within the LLS ice limits on the shores of Loch Linnhe supports a pre-LLS age for original erosion of the platform; at NN 039 657 and NN

015 699 till is present on top of the platform, and Thorp (1984) notes that the LLS ice marginal outwash fans at Kentallen were deposited on the platform.
Fourthly, there is evidence to suggest that pre-existing glacial deposits may not have been moved far down-valley (for example the concentration of glacial drift in upper Glen Resipol, compared to only thin deposits around the terminal moraine).
These observations suggest that the limited duration and thickness of the LLS glacial event meant that it did not result in any major landscape modifications, but merely 'touched up' and abraded existing landforms. As the LLS glaciation was short-lived, this is in accord with the lower end of the range of cited rates of glacial bedrock erosion of 50-5000 mm 1000a $^{-1}$ (Andrews 1975, Coates 1974).

6.2.6 Summary and discussion

Many features of the spatial contrasts in the distribution of glacial erosional and depositional evidence in Lochaber are explained by the influence of tidewater glacier, topographic and geological controls on the dynamics of the LLS ice cap, and the landform legacy of previous glacial episodes. Table 6.2 suggests the relative importance of these factors at the two different spatial scales, based on the degree to which field evidence conforms to theoretical predictions.
Dynamical controls induced by iceberg calving termini may have been particularly important in causing high velocity, erosional regimes towards the western ice margins, and determining stable stillstand locations during deglaciation. Where glacier termini were terrestrial, LLS glacial sedimentation was concentrated at ice margins. Where termini were in tidewater and the glacier snout lay within the fjord system, sedimentation was concentrated in the proglacial submarine basin at the ice maximum. If the tidewater terminus was at the mouth of the fjord, glacial sediment was dispersed and deposited distally in basins on the continental shelf (Fig 6.1). It is likely that similar dynamical controls applied to the Late Devensian ice sheet.

6.3 Holocene geomorphic activity

The recognition of a fresh LLS glacial landscape provides an opportunity to assess the amount of modification by subaerial processes in the last 10,000 years. It is likely that postglacial geomorphic action was most efficient immediately after deglaciation. This is because there would have been large volumes of potentially unstable debris not yet protected from denudation by soil and vegetation cover. Rapid paraglacial slope activity has been recorded in association with retreating Norwegian glaciers (Ballantyne and Benn 1994), and is most likely where slopes are steep, and rainfall abundant, conditions likely to have been widespread in Western Lochaber. In addition, it is thought that the early postglacial period was one of relatively high seismic activity (Holmes 1984, Sissons and Cornish 1982), which may have triggered slope activity.

Fluvial Activity

Fluvial dissection of till and morainic deposits in Western Lochaber is widespread. In some instances material eroded has been resedimented as alluvial terraces infilling basins dammed by rock bars or moraines, or as alluvial cones. Elsewhere it is transported to the sea loch basins. One site allows a quantitative estimation of postglacial fluvial transport. The seismic results clearly show a delta at the mouth of the Cona and Scaddle rivers in Ardgour (Fig 4.4). The Cona and Scaddle catchment is approximately $90km^2$. This delta is approximately 1375m wide and 42m thick, 1400m offshore. This suggests that a total volume of debris in the order of $1.4x10^7$ m^3 has accumulated at this site since it was deglaciated, implying a postglacial fluvial sediment delivery rate of ~ 1,400 m^3a^{-1}. Additional evidence for Holocene fluvial activity is present at the eastern mouth of Glen Tarbert. Here, the river has incised through fluvioglacial outwash to form a series of terraces. This action may be related to falling Holocene sea-levels. A peat raft overlying glacimarine silts and clays and underlying fluvial gravels at Inversanda was radiocarbon dated to 3075±70 conventional radiocarbon years B.P. This confirms that fluvial incision has occurred during the Holocene.

Marine processes

During the Postglacial Marine Transgression (PGMT), and subsequently, erosional notches have been cut into many coastal till, fluvioglacial and moraine deposits, e.g. along the southern shore of Loch Eil. On exposed coastlines storm beach bars were constructed during the PGMT; fine examples exist at Ballachulish and Peanmeanach. These locations reflect exposure to prevailing south-westerly winds, and large volumes of reworkable sediment. On the sea floor, seismic evidence suggests that postglacial rates of sediment accumulation on the loch floor in Loch Linnhe have averaged rates of 1-3cm p.a. Some of this material may represent reworking of older sediments. This contrasts with figures of 1mm p.a. suggested on the basis of palynological evidence in sea floor sediments in Airds Bay in Loch Etive (cited in Edwards et al. 1987), and 0.5-1mm p.a. in Loch Shiel (cited in Boulton et al. 1981).

Weathering

The amount of Holocene weathering of exposed bedrock surfaces is strongly dependent upon lithology. Psammites and course grained igneous intrusions are prone to granular disintegration. On psammites, where more resistant veins of quartzite are present, the differences in the surfaces suggests that 2-5mm of postglacial weathering has occurred. Sections in the coarse grained Strontian granite reveal a rotted surface layer, and bedrock surfaces of this lithology are almost always obscured by a cover of soil and vegetation, which may have developed easily on a layer of weathered regolith. This lithological propensity to rapid granular disintegration may explain the anomalies in the distribution of smoothed rock slabs. The Strontian granite intrusion is surrounded by scoured slabs on schist bedrock, yet there are very few exposed slabs of Strontian granite. An analogous situation is likely on the granitic intrusions flanking the west shore of Loch Linnhe.

Mica schists are subject to weathering and frost cracking along the closely spaced foliation planes. Measurements of slab relief inside LLS limits on these lithologies showed that 6-8mm of postglacial weathering along these planes of weakness is common. Other types of mica schist outcrop have been subject to frost shattering. Many of the fine grained acid and basic intrusions in Western Lochaber have also been subject to small scale frost shattering during the Holocene.

Rates of postglacial weathering and frost shattering have been considerably lower than those prevailing during the LLS, as demonstrated by the clear contrasts in slab surface relief discussed in Section 3.5.2. Slabs affected only by Holocene weathering have experienced on average up to 6mm of surface lowering along lines of weakness, whereas those subjected also to LLS weathering average up to 12mm. Joints greater than 20cm deep are only located in outcrops exposed to LLS frost action, whereas joints in bedrock subject to just Holocene frost action are usually less than 5cm deep.

The good preservation of a glacial landscape in Western Lochaber shows that Holocene geomorphic activity has not been sufficient to dramatically alter the glacial landforms formed during the LLS. Subaerial postglacial weathering has removed striations and friction cracks from exposed bedrock surfaces, but most other LLS landforms remain intact. Fluvial action has been the most significant geomorphic agent as rivers have eroded and transported glacial deposits towards the lochs and sea lochs.

Part II

6.4 Palaeo-environmental and climatic inferences

6.4.1 Introduction

The relationship between glaciers and climate is determined by two parameters; the amount of snowfall and the way mass is lost by ablation. Together these determine the net mass balance profile of a glacier. The Equilibrium Line Altitude (ELA) is located at the altitude where the net mass balance is zero. The mass balance profile of an individual glacier in equilibrium is mainly determined by local climate. Glacier ELAs are thus often considered a critical parameter in the link between ice extent and climatic regime (Sutherland 84b). However, other factors such as accumulation by avalanching and wind-blown snow, ablation by calving and aspect-controlled variations in ablation are frequently of significant local importance. The mass balance profile is an important influence on the response of a glacier to climatic change, but this response is also affected by local topography and valley geometry (Furbish and Andrews 1984). The mass balance profile, ice flow patterns and topography together determine a glacier's areal distribution of ice with altitude, known as the hypsometric curve. As a result of these additional non-climatic influences on glacier hypsometry, caution is necessary when inferring regional ELAs from the reconstructed dimensions of individual former glaciers.

A further problem arises when making palaeoclimatic inferences from estimated ELAs of reconstructed LLS glaciers. This is because the assumption that LLS glaciers were in equilibrium with the climate may be invalid. Modelling experiments by Payne and Sugden (1990a,b) suggest that the maximum extent of the LLS ice cap was not in equilibrium with long term climatic conditions. Rather, it was a fleeting maximum attained before climatic amelioration caused subsequent retreat. As there is a lag time between climatic changes and glacier terminal responses, the maximum position is thus likely to represent a position attained briefly following a truncated advance, rather than an equilibrium position maintained for some time.

In order to estimate palaeo-ELAs the areal distribution with altitude of each reconstructed glacier in Western Lochaber was calculated. This was obtained by firstly dividing the ice cap into separate glacier systems on the basis of basal topography and ice flow direction indicators (Fig 6.5). Secondly, the area within each contour interval was measured. This was achieved by counting 500m squares superimposed on a 1:50,000 map of glacial limits in the case of the main outlet glaciers, and using 250m sided squares and a 1:25,000 map of the small independent glaciers. In the calculations for the large Linnhe glacier a map at a scale of 1:263,000 and a grid with sides of 1.32km was used. The extent of this glacier was estimated from the ice surface contour map presented here, and the work of Sissons (1979a), Thorp (1986) and BGS Sheet 62E (drift edition). This involved estimation of the proportions of ice from Loch Eil and further north and east which flowed south to the Linnhe glacier. The hypsometric curves of each glacier are shown in Fig 6.6.

There are several ways of estimating the ELAs of individual former glaciers using reconstructed ice dimensions. Here, two methods are used. Firstly, ELAs are calculated from the reconstructed glacier hypsometric curves by using estimated Accumulation Area Ratios (AARs) of the former glaciers. Secondly, ELAs are calculated using Sisson's (1974b) method.

6.4.2 ELAs reconstructed from AARs

Various reviews of methods for estimating palaeoglacier ELAs have concluded that a method using estimated former Accumulation Area Ratios (AARs) is often the most accurate (e.g. Meierding 1982, Hawkings 1985, Torsnes et al. 1993). The AAR of a glacier is the area of the accumulation area (above the ELA) at the end of summer as a fraction of the total glacier area. This is likely to be a valid method for inferring palaeo-ELAs of the LLS glaciers in Western Lochaber as the glaciers existed under the same climatic regime, and have similar hypsometric curves (Fig 6.6)(Furbish and Andrews 1984). Measurements of contemporary maritime, land terminating glaciers have shown that glaciers with an AAR of <0.5 are currently retreating, those with an AAR of 0.6-0.7 have stable frontal positions, and those with values >0.8 are advancing. Stable tidewater glaciers, however, can have AARs of ~0.8, and AARs of up to 0.9 when advancing, as their ablation areas are truncated (Meier et al. 1980, Meier and Post 1962, Mann 1986).

It is possible to estimate this calving-induced reduction in surface area by using an empirically derived calving law. Observations suggest that the calving rate of a tidewater glacier is a function of the width-averaged water depth at the ice terminus (Pelto and Warren 1991). The calving flux is the product of the calving rate and the cross-sectional area of the ice terminus. These parameters can be calculated using depth data from Admiralty charts, and an estimated ice cliff height, based on empirical observations of the relationship between cliff height and water depth of contemporary tidewater glaciers (Brown et al. 1982, Warren pers. comm.). Water depth and ice cliff height have been calculated taking into account higher LLS sea-levels, but without adjusting sea floor sediment thicknesses. Using a mass balance curve derived from palaeoecological LLS climatic indicators and an energy balance model (Kerr 1992), the ice surface area at sea level sufficient to allow melting equivalent to the ice mass lost by calving can be estimated. This assumes that there would be no significant change in glacier hypsometry up-glacier if the glacier did not have a

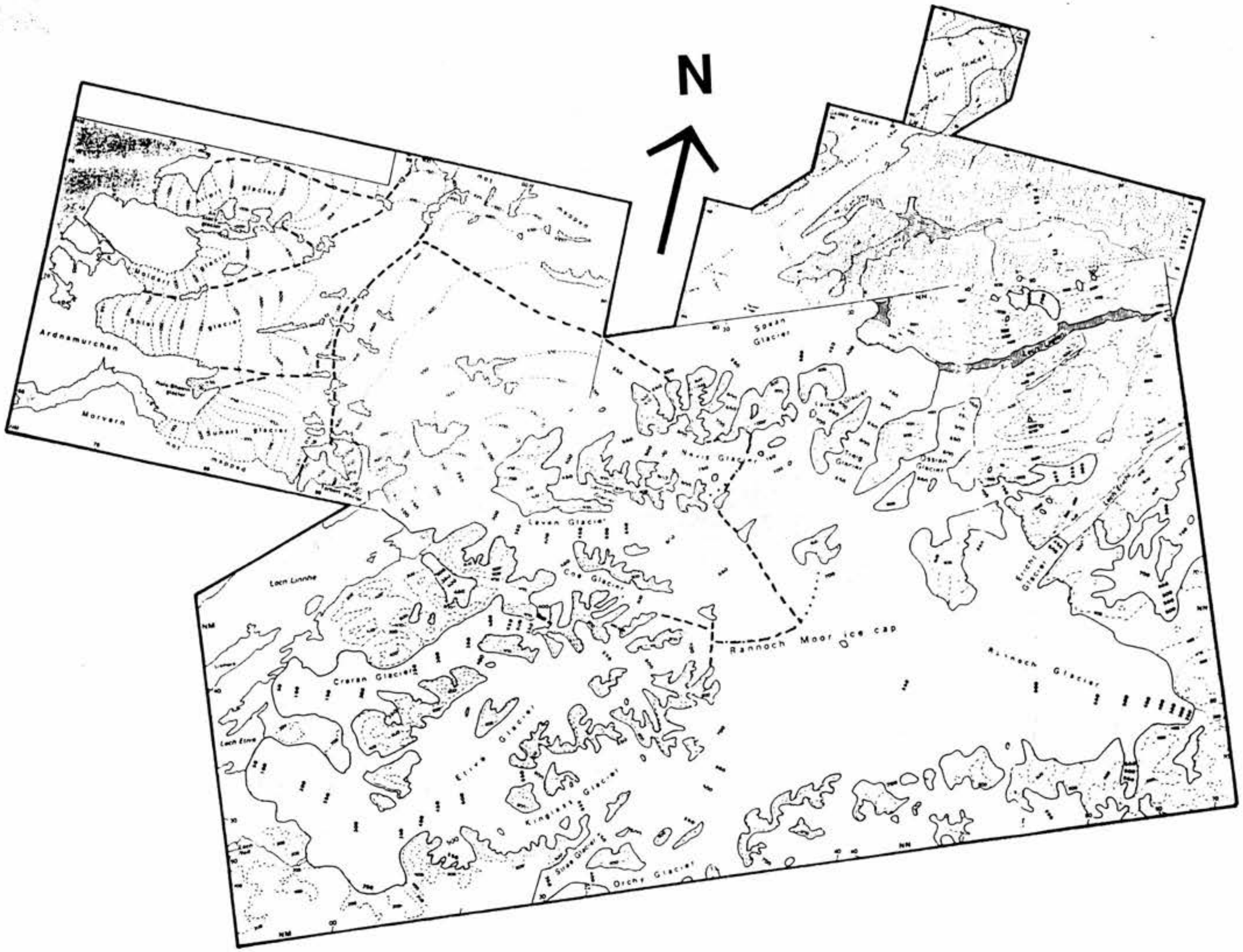
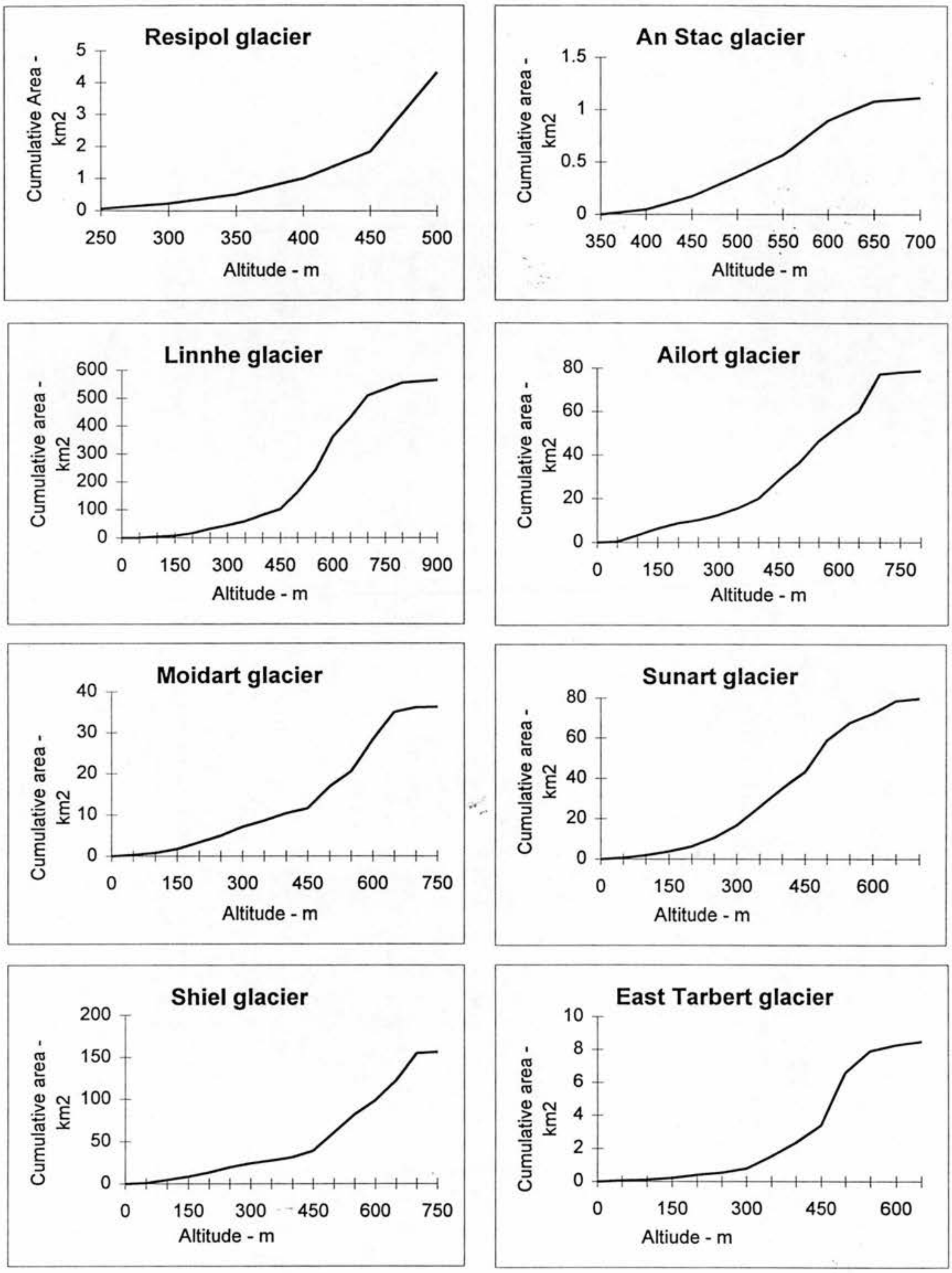


Fig 6.5 LLS ice cap in Lochaber divided into component glaciers. Ice in Eastern Lochaber reproduced from Thorp 1986 and Sissons 1979a.



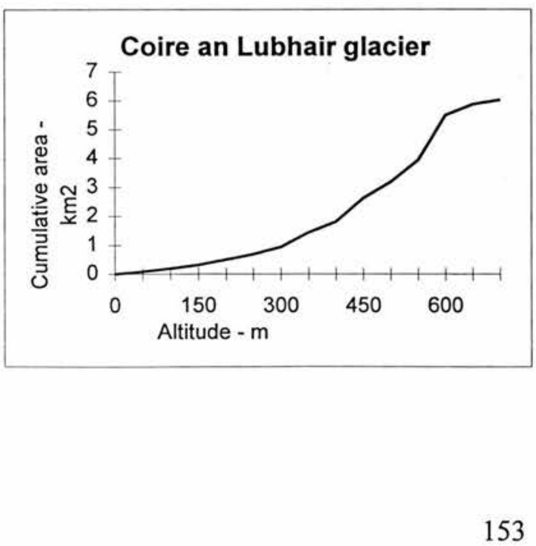


Fig 6.6 Reconstructed glacier hypsometric curves.

calving terminus. The area thus calculated can then be compared with the total glacier area in order to estimate the adjustment necessary to the inferred AAR to account for calving. The results of these calculations are shown in Appendix 7, Table A.7.1. As many of the inputs to these calculations are estimated, and due to the untestable assumptions involved, the results should be interpreted only as first order approximations.

Palaeo-ELAs were obtained by assuming a former AAR for each glacier. These values were different for each glacier, and are shown in Table 6.3. Two alternatives are presented. The first set are based on a equilibrium AAR of 0.65 for terrestrial glaciers, with adjustments made according to Table A.7.1 to account for calving where glaciers terminated in significant water depths. In the second set these AARs are increased by 0.1 to account for the fact that the glaciers may have been advancing and not in equilibrium with the prevailing climate. The use of the AAR method therefore has the advantage over other methods of calculating ELAs that the influence of calving and advancing glacial regimes may be taken into account.

Table 6.3 Reconstructed ELAs for glaciers in Western Lochaber.
*= smaller corrie glaciers.

Glacier	Area - km^2	**1.** AAR	**1.** ELA - m	**2.** AAR	**2.** ELA - m	ELA - m (Sissons)
Ailort	78.9875	0.72	412	0.82	328	481
Moidart	36.2500	0.67	454	0.77	343	469
Shiel	156.5000	0.65	486	0.75	450	513
Sunart	79.7500	0.67	353	0.77	309	393
E. Tarbert	8.4688	0.65	429	0.75	386	436
Coire an Lubhair	6.1313	0.65	418	0.75	359	461
Linnhe	564.69301	0.80	407	0.9	339	547
Mean			**423**		**359**	471
Standard deviation			**38**		**43**	46
Resipol *	4.3125	0.65	430	0.75	404	432
An Stac *	1.1094	0.65	507	0.75	478	535
Total mean			432		377	474
Total standard deviation			43		54	48

6.4.3 ELAs calculated using Sissons' method

The final column of Table 6.3 shows ELAs calculated using the method of Sissons (1974b). This method assumes a linear mass balance gradient, and that the glaciers are in equilibrium with the prevailing climate. Since mass balance curves reconstructed for Western Scotland are not linear (Kerr 1992), since mass lost by iceberg calving will exacerbate this non-linearity, and since the LLS maximum is unlikely to represent a steady state equilibrium condition (Payne and Sugden 1990a,b), this method has limitations. However, since all other empirically based reconstructions of LLS ELAs in Britain have used Sissons' method, it is included here for comparative purposes.

6.4.4 Discussion

The reconstructed ELAs of individual glaciers in Western Lochaber have a mean of 360-480m with a standard deviation of 38-54m. The results from different glaciers are thus fairly consistent, lending support to the idea that they may be a reasonable approximation to a regional, climatic ELA. Theoretical considerations discussed here suggest that the values derived using Alternative 2 are most likely to represent the true values, and a regional ELA of ~360m is suggested. This ignores the values obtained from the two smaller corrie glaciers since the ELAs of such glaciers are less likely to approximate to regional ELAs due to the large potential influences of aspect, shadow and wind-blown snow on accumulation and ablation rates (Meierding 1982).

There are good reasons which may explain why some glaciers show ELAs which deviate from the regional mean value. Although the Shiel glacier did not have a calving terminus, during much of the advance period it had a terminus in water which is at present up to 150m deep. It is therefore likely that advance was impeded by high calving fluxes, and thus may have lagged in time compared to adjacent non-calving glaciers. This would have the effect of causing the reconstructed ELA to be higher than the actual ELA, which may explain why the ELA derived for this glacier is the highest. The value obtained for the Sunart glacier may be anomalous due to inaccurately reconstructed glacier hypsometry as neither the southern lateral margin or the terminal position are well constrained. The ELA reconstructed for the An Stac glacier is higher than the others. This may be due to topographic controls as the terminal moraine of this glacier is superimposed on a bedrock bar. Alternatively, it is possible that this moraine represents a retreat stage, and that at it's maximum the glacier was confluent with the Ailort glacier in the trough below.

Comparison of these ELA altitudes with those obtained by Sissons' methods show that, for land terminating glaciers, both methods often produce broadly comparable results. However, Sissons' method overestimates palaeo-ELAs of tidewater and advancing glaciers. The ELAs derived here are comparable with those calculated from the dimensions of LLS glaciers elsewhere in this part of Scotland using Sissons' (1974b) method. Thorp (1984) estimated paleao-ELAs of 419m and 432m for the Etive and Creran glaciers respectively, and a mean value of 319m has been derived for the LLS glaciers on the Isle of Skye (Ballantyne 1989). Sissons (1979b) reconstructed a regional ELA of ~250m for the Isle of Mull, and Wain-Hobson (1981) suggests values of 290-390m for glaciers in Morvern. However, the dimensions of these glaciers in Mull and Morvern may have been underestimated (Benn and Evans 1993).

Previous analyses of reconstructed ELAs for LLS glaciers in Scotland have identified a marked west - east increase in ELA altitudes (Sissons 1974a,1979b). This is thought to be due to a strong west - east precipitation gradient (Sissons 1980). The palaeo-ELAs reconstructed here provide additional support for this pattern, as they confirm the lower ELAs in Western Scotland. Re-investigation of this national pattern taking into consideration advancing and calving glacial regimes, and recent suggestions of more extensive and thicker ice in certain locations (e.g. Benn and Evans 1993, Thorp 1986, Ballantyne 1989, Lambeck 1993, Bennett and Boulton 1993a,b) may shed further light on this pattern.

6.4.5 Climatic inferences

If the mean ELA derived here represents the regional, or climatic, ELA for Western Lochaber during the LLS, some inferences may be made about prevailing climate. Since the climate is likely to have varied during the stadial, since the glacial response time to climatic change and the influences of bedrock topography, ice flow patterns, and local factors influencing rates of accumulation and ablation on individual glacier ELAs are unknown, such inferences must be made with extreme caution. In addition, there is no unique set of climatic parameters which produce a given ice volume. Computer modelling experiments have used different climatic inputs to 'grow' the LLS ice cap. Payne and Sugden (1990a,b) used a model which generated an ice mass in good agreement with empirical evidence using as input temperature and precipitation curves which peaked at a -8^{o}C July temperature depression (equivalent MAAT depression = ~ -21^{o}C) and ~60% of contemporary precipitation. In contrast, the model described by Kerr (1992) 'grows' a similar ice cap, over a similar timespan, with a MAAT depression of -7^{o}C and 50 % of contemporary precipitation. Both these models use a topographic grid of a 5km scale. Experiments using different inputs to the high resolution, 1km grid numerical ice sheet model discussed in Section 5.2.3 suggest that only certain scenarios could have produced the reconstructed LLS ice mass in Western Lochaber. The model suggests that ice does not accumulate in Western Lochaber unless the local ELA drops to approximately 300m, somewhat lower than that calculated on the basis of reconstructed ice dimensions. Western Lochaber is covered by ice flowing from the Ben Nevis area if the ELA is 575 - 300m.

Extrapolation of the relationship between ELA altitudes and climatic parameters in Western Scotland presented by Payne (1988) suggests that the regional ELA of ~360-420m calculated here on the basis of reconstructed ice dimensions may be associated with climatic regimes characterised by climatic scenarios ranging from a mean July temperature depressions of -7.2^{o}C to -8.75^{o}C (equivalent MAAT depressions ~ -19^{o}C to -23^{o}C) with 2.4ma^{-1}(100% of present values) to 1.2ma^{-1} (50%) precipitation respectively. According to this relationship, the ELA of 300m suggested by the high resolution modelling experiments is associated with a mean July temperature depression of -8^{o}C if precipitation was 100% that of present. The relationship between the same parameters described in Kerr (1992) suggests the reconstructed ELA could be produced by MAAT depressions and percentages of contemporary precipitation of between -7.5^{o}C with 100% precipitation to -12^{o}C with 50% precipitation.

Compilation of these climatic scenarios produced from different means of inferring climate from ice dimensions suggests a broad envelope of climatic possibilities between MAAT depressions of -7.5^{o}C with 100% precipitation, and MAAT depressions of -23^{o}C with 50% precipitation. Independent periglacial and palaeoecological evidence suggests that the MAAT depression was at least ~ -17^{o}C during the LLS (Ballantyne and Harris 1994), so scenarios closer to the colder and drier end of the range may be more likely. In view of the uncertainties involved in linking ice extent to climatic regime, these inferences must be regarded as tentative.

Chapter 7 - Conclusions

7.1 Aim

This chapter summarises the main conclusions and wider implications.

7.2 Summary

Geomorphological evidence shows clear ice limits in some troughs in Western Lochaber and limiting constraints for the positions of ice limits in other troughs. Geomorphological mapping also gives evidence for ice flow directions and shows a clear contrast in the terrestrial distribution of glacial erosional and depositional features in Western Lochaber. In particular, erosional landforms are widespread in the west, and glacial till and moraines are much more prevalent in the north east of Western Lochaber. It has been demonstrated that glacial trimlines provide a reliable indication of palaeo-ice surfaces. Out of 110 determinations of trimline altitudes, 64% show firm evidence of ice margins, and all produce glaciologically consistent values. Evidence for palaeo-ice margins is augmented by the results of a seismic survey in Loch Linnhe. The clarity of the seismic record in many parts of this loch enables the identification of a clear ice limit south of the Corran narrows. The location of this limit corresponds exactly with terrestrial evidence for an ice limit on the eastern loch shore mapped by Thorp (1986).

Combining the geomorphological, trimline and seismic evidence allows the reconstruction of an ice cap in Western Lochaber. These three lines of evidence are all independent, yet consistent. This ice cap was up to 650m thick, and extended into the sea lochs. The ice surface was highest in the north east of Western Lochaber and around the main mountain ridges. Deglaciation involved retreat back from the coasts towards the mountains, punctuated by stillstands at topographic pinning points in the sea lochs. The most important non-climatic controls on palaeo-ice dynamics are likely to have resulted from iceberg calving termini in the sea lochs. Iceberg calving contributed to high velocity ice flow in these troughs, and under these circumstances glacial debris was efficiently evacuated to the ice snouts and deposited on the sea bed.

? reached up to 650m

The dimensions of the Western Lochaber ice cap suggest a mean glacier ELA of ~ 360m during the LLS. Different formulations of the relationship between climate and ice dimensions suggest that this ice mass may have been produced by a broad range of climatic possibilities from MAAT depressions of -7.5°C with 100% precipitation to MAAT depressions of -23°C with 50% precipitation. The latter end of this range is consistent with independent periglacial and palaeoecological evidence for LLS climates established elsewhere.

7.3 Overview

Detailed field studies using several different types of field evidence have highlighted the role of tidewater glaciers as controls on LLS ice cap dynamics in Western Lochaber. Topographic influences on calving rates determined stable positions for ice maxima and retreat stillstands and may have influenced rates of advance and retreat. In particular, the Linnhe calving terminus may have been particularly important. This is suggested both by estimations of palaeocalving fluxes (Section 6.3.2) and by the results of the high resolution numerical ice sheet modelling experiments. These produce an ice mass reconstruction different to that produced on the basis of field evidence if iceberg calving in the Linnhe trough is not taken into account. The results suggest that the Linnhe glacier was such an important ice sheet outlet in terms of ablation that its effect was to bisect an ice cap to produce two ice domes. Any reconstruction of controls on LLS ice cap morphology and dynamics elsewhere in Western Scotland may also need to consider the influence of tidewater glaciers.

Tidewater glacier behaviour has also exerted controls on the locations of glacial erosional landforms and deposits. Where termini are terrestrial, till and moraines are widespread. Where termini are offshore deposition is concentrated at the ice front in the form of outwash fans around the lochs or as submarine glacimarine deposits. This has important implications for the use of field evidence to reconstruct the dimensions and dynamics of Quaternary ice sheets in Scotland and other coastal locations. In particular, reconstructing palaeoice margins on the basis of terrestrial moraines and till deposits alone may be problematic where glaciers had calving termini. In these situations a wider variety of evidence can be linked together. Terrestrial proglacial outwash fans with large volumes or ice contact slopes may indicate stillstand positions around the loch shores. Seismic surveys and offshore boreholes may provide evidence for marine limits. Since glacial erosion rather than deposition may predominate throughout the lengths of former tidewater glaciers, trimlines may be the most appropriate means of reconstructing former ice limits in the mid and upper reaches of such palaeoglaciers. This is because the clarity of trimline evidence is enhanced by efficient glacial scour and an absence of superficial deposits.

It is likely that the extent and fluctuations of calving glaciers do not directly reflect climatic controls. Reconstructed palaeocalving fluxes suggest that, for deep and wide troughs such as Loch Linnhe, glacier hypsometry may be significantly influenced by high calving fluxes.

Reconstructions of palaeo-ELAs based on hypsometry of calving glaciers are thus problematic and the effect of truncated ablation areas must be taken into account. Due to this non-climatic influence on ablation rates, the ELAs of calving glaciers are less likely to be representative of regional, climatic ELAs than the those of land terminating glaciers. LLS palaeoclimatic reconstructions inferred from glacier dimensions are thus likely to be most valid where termini were not calving.

7.4 Future work

In order to build on the work in this thesis it would be useful to date rock surfaces above and below trimlines using cosmogenic surface exposure age dating techniques. This would enable further evaluation of the validity of trimlines as former ice surface indicators, and provide additional constraints on the age of the ice cap reconstructed here. Secondly, offshore borehole and further seismic surveys may provide improved evidence for ice limits in the sea lochs. The clear seismic records obtained from Loch Linnhe provide excellent evidence for a former ice limit. Further offshore investigations in Western Scotland would determine whether similar clear evidence exists in other sea lochs. If so, this could provide additional constraints on ice limits to be tied in with terrestrial evidence, and provide further evidence for controls tidewater termini may exert on the location of glacigenic sediment. Comparisons of palaeoglaciers which had terrestrial and tidewater termini with the distribution of glacial deposits and erosional features over a wider area of Scotland would shed further light on the controls that the dynamics of tidewater glaciers may exert on the glacial geomorphological record.

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Appendices

Contents

Appendix 1. Clast shape and roundness data

Table A.1 Aggregate clast shape and roundness indices of samples from fluvioglacial outwash, rivers and beaches in Western Lochaber.

	RA index	C40 index	RR index
Beaches			
Eil Beach	6	28	20
Forsay Beach	0	24	42
Rivers			
Finnan River	6	26	32
Suileag 1	48	8	0
Suileag 2	10	26	14
Suileag 3	0	28	14
Suileag 4	0	8	16
Basal tills			
Doilet Basal Till	6	22	2
Druim Beag grey	44	52	0
Meltout tills			
Tarbert Meltout Till	96	4	0
Fionnlaighe	58	12	0
Outwash			
Inverscaddle Outwash	60	34	2
Corualasnacon	18	16	0
Aladale	6	32	2
Annat south	6	18	4
Annat north	12	28	10
Corran east	8	0	10
Corran west	0	16	16
Arivegaig	0	12	34
Acharacle dump	2	20	24
Shiel Bridge 1	2	8	18
Shiel Bridge 2	0	20	32
Shiel Bridge 3	0	12	10
Dalnabreck	2	20	24
Dalelia	0	16	18

Clast angularity indices

RA index = % of clasts in VA and A categories (Benn and Ballantyne 1994)

RR index = % of clasts in WR and R categories

Clast shape index (Benn and Ballantyne 1994)

C40 index = % of clasts with C:A axial ratio of ≤ 0.4

High C40 indices are associated with unmodified frost shattered debris

Low C40 indices are associated with actively transported clasts

Appendix 2 Clast lithology counts

Location	Glenfinnan Col	Glenfinnan	Glen Dubhlighe	Glen Fionnlighe	Glen Suileag	Glen Moidart	River Ailort	Shiel Bridge	Loch Eil moraine	Dalnabreck	Glen Hurich section 5
Psammite	32	32	24	24	57	18	24	26	28	28	30
Pelite	4	4	8	20	4	10	20	3	5	6	10
Micaschist	6	4	8	0	2	30	10	3	3	10	6
Quartzite	18	20	8	22	17	6	12	20	18	20	20
Schist/ Granite mix	0	0	6	8	0	0	0	0	0	0	0
Banded quartzite + psammite	4	8	10	4	0	12	12	0	0	0	8
Granitic gneiss	22	14	20	4	0	8	12	20	10	10	14
Scaddle intrusion (Gabbro)	0	0	0	0	0	0	0	0	0	2	0
Pegmatite	0	6	2	4	6	10	0	8	12	4	4
Strontian granite	0	0	0	0	0	0	0	0	0	0	0
Appinite	0	0	0	0	0	2	0	2	0	2	0
Other intrusions											
Fine acid	8	8	14	14	13	4	2	18	23	18	6
Black + white	6	0	0	0	2	0	0	0	0	0	0
Pink + black + white	0	2	0	0	0	0	4	0	0	0	2
Dolerite	0	2	0	0	0	0	0	0	0	0	0
Agglomerate	0	0	0	0	0	0	0	0	0	0	0
Rhyolite	0	0	0	0	0	0	0	0	0	0	0
Lamprophyre	0	0	0	0	0	0	0	0	0	0	0

Location	Polloch bridge	Inversanda beach	Mid Glen Tarbert	Annat	Glen Gour	Resipol valley	Corran	Glen Scaddle	Cona Glen	Callop	Stonecreggan Glen
Psammite	37	13	9	53	25	27	27	43	22	38	44
Pelite	4	4	4	2	5	8	6	4	7	4	2
Micaschist	16	4	0	0	2	21	0	2	4	4	6
Quartzite	8	17	9	18	22	14	23	12	27	7	24
Schist/ Granite mix	0	2	6	2	3	0	2	2	9	0	0
Banded quartzite + psammite	8	0	6	0	0	0	0	2	2	0	0
Granitic gneiss	14	40	57	6	0	20	6	0	9	19	2
Scaddle intrusion (Gabbro)	0	0	0	0	6	0	2	22	4	0	0
Pegmatite	4	0	0	8	2	0	4	2	7	7	0
Strontian granite	0	0	0	0	5	4	0	0	0	0	2
Appinite	0	0	0	0	0	0	0	0			
Other intrusions											
Fine acid	10	21	9	8	27	2	21	7	7	21	20
Black + white	0	0	0	0	0	0	4	0	2	0	0
Pink + black + white	0	0	2	0	0	0	0	4	0	0	0
Dolerite	0	0	4	2	0	0	0	0	0	0	0
Agglomerate	0	0	0	0	0	0	0	0	0	0	0
Rhyolite	0	0	0	0	3	0	5	0	0	0	0
Lamprophyre	0	0	0	0	0	4	0	0	0	0	0

Glen	Lochain Dubh	Cloiche Sgoilte	Mhic Phail	Upper Scaddle	Caol	E. Linnhe	Upper Callop	North Gour	Strontian
Psammite	34	38	44	37	41	40	48	8	2
Pelite	0	0	0	4	4	0	2	0	0
Mica schist	13	2	0	4	0	0	4	2	0
Quartzite	11	8	10	16	13	19	4	30	8
striped Ps + Q	0	0	0	0	0	0	0	0	2
Biotite schist	9	2	4	4	0	0	0	0	0
Granitic gneiss	17	24	10	10	0	0	30	0	54
Scaddle intrusion	4	0	4	2	0	0	0	22	0
schist + intrusion	2	0	0	2	0	0	2	0	0
Pegmatite	2	12	16	6	0	4	4	2	6
Strontian granite	0	0	0	0	0	0	0	0	12
Appinite	0	0	0	0	0	0	0	0	0
Other intrusions									
coarse acid	2	0	2	0	7	8+2	0	0	0
medium acid	4	12	0	12	0	15	4	16	0
fine acid	2	0	0	0	4	2	0	2	12
Diorite	0	0	0	0	2	0	0	14	0
microdiorite	0	0	0	0	0	0	0	0	0
andesite	0	0	0	0	11	6	0	0	0
lava	0	0	0	0	2	2	0	0	0
medium basic	0	0	0	0	0	0	0	2	0
fine basic	0	0	0	0	0	2	2	0	4

Appendix 3 Raised marine features

Table A.3.1 Raised marine evidence for former sea levels
* measured using E.D.M., @,# measured using Abney level

Loch	Location	Grid Reference	Feature	Measured Altitude - *= m O.D. @= m above m.s.l. #=m above MHWS.	Assumed altitude of formation	Implied mean sea level at time of formation - m O.D.
Nan Uamph	Loch Beag	NM 733 833	Raised beach	@8 - 9.6	MHWS	6.1 - 7.7
	Druimindarroch	NM 686 845	Raised beach	@7.11	MHWS	5.21
	Arisaig House	NM 695 850	Raised beach	30-40m O.D. (O.S. map)	MHWS	~30-40
	Prince Charlie's Cave	NM 692 843	Beach bar	#5.79	MHWS +2m	4.04
Ailort	Samalaman	NM 661 773	Raised beach	~30m O.D (O.S. map)	MHWS	~28
	West Glenuig	NM 668 777	Summit of raised beach	@16.0	MHWS	14.24
	Forsay	NM 688 780	Back of raised beach	#14.22 above beach bar	MHWS +1m	14.48
	Peanmeanach	NM713 805	Storm beach bar	#9.48	MHWS + 2.5m	7.24
	Laggan	NM 719 799	Raised beach	#7.11	MHWS	7.37
	Alisary	NM 742 795	Raised delta	#~7.9 (no clear breaks in slope)	MHWS	~8.16
	S. Alisary	NM 739 790	Raised delta	#~7.9 (no clear breaks in slope)	MHWS	~8.16
	Coopers Knowe	NM 712 791	Raised beach	@6.4 - 8	MHWS	4.84 - 6.44
	Coopers Knowe N	NM 714 792	Beach notch	@4.8	MHWS	3.24
Moidart	Shona Beag	NM 669 739	Raised marine sands?	# up to 9.33	< MHWS	>9.61
Shiel / Moidart	Torr Mor	NM 665 713	Raised beach	~30m O.D. (O.S. map)	MHWS	~30

Loch	Location	Grid Reference	Feature	Measured Altitude - *= m O.D. @= m above m.s.l. #=m above MHWS.	Assumed altitude of formation	Implied mean sea level at time of formation - m O.D.
Shiel	Arivegaig (river bank)	NM 655 683	Raised marine sands?	*5.58	<MHWS	>3.50
	Arivegaig (quarry)	NM 679 660	Subaerial outwash gravels?	*6.64	>MHWS	<4.56
	Shiel Bridge	NM 678 691	Subaerial outwash gravels?	*6.87	>MHWS	<4.79
	Moss	NM 680684	Beach notch	# 3.16 above a high Loch Shiel (~4.5m O.D.)	>/= MHWS	</= 7.96
	Claish Moss - Rubha Daraich	NM 725 685	Subaerial outwash gravels?	*6.63	>MHWS	<4.55
	Langal	NM 709 696	Raised beach?	*8.86	MHWS	6.78
	Langal	NM 709 696	Beach sediments	# =/> 1.58 above a high Loch Shiel (~4.5m O.D.)	</= MHWS	>/= 4.02
	Gorstanvorran	NM 806 718	Raised beach?	#3.4 above Loch Shiel	MHWS	5.32
Sunart	Resipole	NM 722 644	Fluvial terrace graded to higher sea level	#9.8 above beach bar	>MHWS ?	<11.2
	Resipol	NM 723 640	Raised beach	#7.9 above beach bar	MHWS	9.3
	Laudale	NM 756 599	Low beach bar	*6.56	MHWS + 0.5m	4.06
	Strontian	NM 817 617	At or above lower limit of undisturbed outwash	#~10.82 m OD (from spot height)	>MHWS	<~11.46
	Carnoch W.	NM 839 606	Low beach bar?	*9.12	MHWS +1m	6.12
	Carnoch E.	NM 843 608	Low beach bar?	*10.31	MHWS +1m	7.31

Loch	Location	Grid Reference	Feature	Measured Altitude - *= m O.D. @= m above m.s.l. #=m above MHWS.	Assumed altitude of formation	Implied mean sea level at time of formation - m O.D.
Linnhe	Kentallen	NM 992 577	Beach bar	#12.64	MHWS +2.5m	11.21
	Kentallen South fan	NM 978 571	Notch in deposits	@5.37	MHWS	4.17
	Kentallen North fan	NN 003 579	Notch in deposits	@6.10	MHWS	4.9
	Kentallen North fan	NN 003 579	Notch in deposits	@3.00	MHWS	1.8
	Inversanda	NM 943 591	Raised beach	@11.05	MHWS	9.85
	Inversanda	NM 944 591	Notch	@5.0	MHWS	3.8
	Gearradh S.	NM 954 602	Beach bar	*14.22	MHWS +2m	10.45
	Gearradh N.	NM 962 609	Low beach bar	*13.87	MHWS +1.5m	10.60
	S. Sallachan	NM 968 616	Notch in deposits	#12.64 above beach bar	MHWS	14.21
	North Ballachulish	NN 053 605	Storm beach bar	*14.57-15.38	MHWS +2.5m	10.22-11.03
	South Ballachulish	NN 059 593	Raised beach	*14.41	MHWS	12.56
	Onich	NN 021 616	Raised beach bar	*14.32	MHWS +2m	10.97
	Clovulin	NN 005 635	Notch	*9.42	MHWS	7.57
	Clovulin	NN 005 633	Low beach bar	*6.84	MHWS +1m	3.99
	Clovulin	NN 005 632	Notch	*5.30	MHWS	3.45
	Corran	NN 008 645	Notch	#10.65	MHWS	11.14
	Trislaig	NN 091 744	Low beach bar	@6.9	MHWS +1m	4.92
Eil	Achaphubuill	NN 079 760	Notch	#9.48	MHWS	10.31
	W. Eil narrows	NN 018 769	Raised beach	*13.48	MHWS	11.88
	Blaich	NN 027 769	Notch	*13.22	MHWS	11.82
	Blaich	NN 044 771	Notch - good	#11.30	MHWS	12.33
	Blaich	NN 031 769	2 Notches - good	#11.06	MHWS	12.09
	Blaich	NN 028 769	Notch	*8.40	MHWS	7.00
	Blaich	NN 028 769	Notch	*7.78	MHWS	6.38
	W. Blaich	NN 026 769	Notch	#9.48	MHWS	10.51
	Duisky	NN 003 775	Notch	#11.06	MHWS	12.09
	E. Garvan	NM 983 776	Notch	#6.50	MHWS	7.53
Leven	Coalasnacon	NN 139 612	Low beach bar	*7.77	MHWS +1m	5.03
	Coalasnacon North shore E	NN 146 618	Raised beach	*12.66	MHWS	10.92
	Coalasnacon North Shore W	NN 135 614	Raised beach	*8.11	MHWS	6.37

Table A.3.2 Altitudes of Mean sea level (MSL) and Mean High Water Springs (MHWS) above O.D. (Newlyn), calculated from tidal ranges and local chart datums.

Location	MHWS - m above local datum	M.S L - m above local datum	Local Datum - m O.D.	MSL - m O.D.	MHWS - m O.D.
Mallaig	5.0	2.83	-2.62	0.21 #	2.38
Arisaig			L.A.T. *	0.3	2.4
Outer Loch Ailort (estimated)			L.A.T. *	~0.26	~2.28
Mid Loch Ailort (estimated)				~0.26	~2.08
Mouth of Loch Moidart	4.8		L.A.T. *	~0.28	~2.18
Kentra				0.3	2.38
Salen - Loch Sunart	4.6	2.6	-2.20	0.4	2.4
Port Appin	4.2	2.5	-1.95	0.55	2.25
Kentallen / Inversanda (estimated)				~0.57	~2.34
Corran	4.4	2.55	-1.96	0.59	2.44
Ballachulish (EDM altitude)					2.58 (beach ridge)
Mid Loch Leven (estimated)				~0.56	~2.30
Corpach	4.1	2.50	-1.98	0.52	2.12
Mid Loch Eil (estimated)				~0.50	~1.90
Mid Loch Eil (EDM altitude)					2.43 (highest seaweed line)

*= Lowest Astronomical Tide - unspecified altitude
#= obtained from Admiralty tidal prediction plot.
Sources: Admiralty charts, Shennan et al. 1994a, Shennan 1994.

Appendix 4 Trimline evidence

Table A.4.1 Results of trimline mapping on slopes and spurs

Glen	Hillslope	Altitude m	Fresh glacial evidence below	No glacial evidence above	Boulder concentrations at trimline	No clear periglacial evidence below	Widespread periglacial evidence above	Clarity	Comments	Trimline no.
Ailort	Creag Bhan S	440	✓	×	✓	✓	✓	firm		T1
	Seann Chruach NE	470	✓	✓	✓	✓	✓	firm		T2
	An Stac E	530	✓	✓	✓	✓	✓	firm	mica schist	T3
	Diollaid Bheag NE	525	✓	✓	✓	✓	✓	firm		T4
	Diollaid Mhor SE	670	✓below 550m	✓	×	×	✓	probable	steep	T5
	Beinn Mhic Cedidh N	670	✓	×	×	✓	✓	firm		T6
	Glas-charn S	600	✓	✓	×	✓	✓	firm	mica schist, steep face outcrops	T7
	Sgurr an Utha W	</= 760	✓	×	×	✓	✓	probable	scoured slabs at all altitudes	T8
	Beinn Odhar Mhor NW	700-730	✓	×	✓	✓	✓	firm		T9
Moidart	Brunery Hill S	375-390	✓	×	✓	×	✓	probable	slopes obscured by bracken	T10
	Brunery Hill SE	>/= 390	✓	×	×	×	✓	probable	mica schist	T11
	Ceann Loch Uachdrach N	230	✓	×	✓	✓	×	probable		T12

Glen	Hillslope	Altitude m	Fresh glacial evidence below	No glacial evidence above	Boulder concentrations at trimline	No clear periglacial evidence below	Widespread periglacial evidence above	Clarity	Comments	Trimline no.
	Creag nan Lochan NE	>460	✓	×	×	✓	×	indistinct		T13
	Beinn Gaire SW	480	✓	×	×	✓	✓	firm	few outcrops, free faces	T14
	Sgurr Dhomhuill Beag E	520-620	×	✓	×	×	✓	probable	steep, mica schist	T15
	Beinn Gaire N	580	✓	✓	✓	✓	✓	firm	steep, free faces, closely foliated schist	T16
	Sgurr na ba Glaise S	580	✓	✓	✓	✓	✓	firm		T17
	An t-slat- bheinn E	630	✓	×	✓?	✓	✓	firm		T18
Aladale	Beinn Mhic Cedidh SW	460-680	✓	✓	×	×	✓	indistinct	steep, mica schist, shattered free face outcrops	T19
	Beinn a' Chaorainn W	(500-) 660	✓	×	×	×	✓	probable	mica schist	T20
	Beinn Odhar Bheag W	650-780	✓	×	×	×	✓	indistinct	mixed evidence at high altitudes	T21
Shiel	Resipol NE	480	✓	×	×	✓	✓	firm		T22

Glen	Hillslope	Altitude m	Fresh glacial evidence below	No glacial evidence above	Boulder concentrations at trimline	No clear periglacial evidence below	Widespread periglacial evidence above	Clarity	Comments	Trimline no.
	Sgurr an Easain SW	520	✓	✓	×	✓	✓	firm		T23
	Sgurr an Easain N	600	✓	✓	✓	×	✓	firm		T24
	Sgurr an Tarmachain N	(580-) 670	✓ up to 580m	✓	×	×	✓	probable	free face outcrops	T25
	Beinn a' Chaorainn SE	600-640	×	✓	×	×	✓	probable	steep, mica schist, free face outcrops	T26
	Meall nan Creag Leac W	640-670	✓ up to 530m	✓	✓	×	✓	probable	free face outcrops	T27
	Sgurr Ghiubhsachain N	660	✓	×	×	✓	✓	firm		T28
	Beinn Odhar Mhor E	670-720	✓	×	×	✓	✓	firm		T29
Polloch / Hurich	Sgurr an Tarmachain S	>580	✓	✓	×	×	✓	probable	closely foliated schist, gentle slope	T30
	pt. 770 SW	640	✓	×	×	✓	✓	firm		T31
	Beinn Mheadhoin S	660	✓	✓	×	✓	✓	firm		T32
	Carn na Nathrach S	670	✓	✓	✓	×	✓	firm		T33
	Pt 803 N	670	✓	×	×	✓	✓	firm		T34
	Druim Garbh N	645	✓	×	✓	✓	✓	firm		T35

Glen	Hillslope	Altitude m	Fresh glacial evidence below	No glacial evidence above	Boulder concentrations at trimline	No clear periglacial evidence below	Widespread periglacial evidence above	Clarity	Comments	Trimline no.
Sunart/ Strontian	Garbh Choire E	260	✓	×	✓	✓	✓	probable	weathering difference only	T36
	Beinn a' Chaorainn N	460	✓	✓	×	✓	✓	firm		T37
	Sgurr nan Cnamh NW	(440-) 660	✓ up to 440m	×	×	✓	✓	probable	gentle slope	T38
	Sgurr nan Cnamh N	>/=640	✓	×	×	✓	✓	probable		T39
	Sgurr na h'Ighinn W	650	✓	×	✓	✓	✓	firm	ridge top	T40
	Sgurr na h'Ighinn SE	610-720	✓	✓	×	×	✓	indistinct	steep	T41
	Sgurr a'Chaorain NW	(>/=) 620	✓	×	×	×	✓	probable	steep, massive rock	T42
	Sgurr a'Chaorain SE	600-700	✓	×	✓	×	✓	probable		T43
Tarbert	Creach Bheinn N	490	✓	×	✓	×	✓	firm	ridge top	T44
	Meall a' Chulinn W	500	✓	×	✓	×	✓	firm	gentle slope	T45
	Meall a' Chulinn S	460-500	✓ up to 380m	✓	×(present below)	✓	✓	firm	steep, free face crags around trimline	T46
	Meall a' Bhainache NW	500-540	✓	×	×	✓	✓	firm		T47

Glen	Hillslope	Altitude m	Fresh glacial evidence below	No glacial evidence above	Boulder concentrations at trimline	No clear periglacial evidence below	Widespread periglacial evidence above	Clarity	Comments	Trimline no.
	Meall a'Choirein Luachraich N	440-460	✓	✗	✓	✓	✓	firm	massive gneiss	T48
	Garbh Bheinn S	500	✓ up to 350m	✓	✗	✗	✓	probable	few outcrops	T49
	Meall a' Bhraghaid N	370	✓	✓	✓	✓	✓	firm		T50
	Garbh Bheinn SE	300	✓	✗	✓	✓	✗	probable	weathering difference only	T51
Coire an Lubhair	Druim an Lubhair S	0-130	✓	✗	✗	✓	✓	indistinct	vague weathering difference	T52
	Druim an Lubhair W	400	✓	✓	✓	✓	✓	firm		T53
	Garbh Bheinn E	450	✓	✓	✗	✓	✓	firm		T54
	Sgorr Mhic Eacharna S	570	✓	✓	✓	✓	✓	firm		T55
	Garbh-Bheinn NE	560	✓	✗	✓	✓	✓	firm		T56
Gour	Meall Dearg Coire nam Muc	190-515	✓	✓	✗	✓	✓	indistinct		T57
	Beinn Leamhain NE	</=300	✓	✗	✗	✗	✓	indistinct	geological change	T58
	Beinn Leamhain NW	350	✓	✗	✗	✗	✓	probable		T59
	Sgorr Mhic Eacharna N	(>/=) 500	✓	✓	✗	✗	✓	probable	geological control?	T60
	Beinn na h-Uamha E	470-540	✓	✗	✗	✓	✓	probable		T61

Glen	Hillslope	Altitude m	Fresh glacial evidence below	No glacial evidence above	Boulder concentrations at trimline	No clear periglacial evidence below	Widespread periglacial evidence above	Clarity	Comments	Trimline no.
	Beinn na h-Uamha S	570	✓	✗	✗	✗	✓	probable		T62
	Beinn Bheag N	570-630	✓	✓	✓	✓	✓	firm	topographic change	T63
Linnhe	Meall Breac E	0-500	✓	✗	✗	✗	✓	indistinct	only steep crags exposed	T64
	Beinn na Cille SE	250-400	✓	✓	✗	✓	✓	probable	three geological changes	T65
	Stob Coire a' Chearcaill E	460-500	✓	✓	✗	✗	✓	firm	vegetated	T66
Scaddle	Beinn na Cille NE	460-490	✓	✗	✗	✓	✓	firm		T67
	Beinn na Cille NW	485	✓	✗	✗	✓	✓	firm		T68
	Meall an Tairbh N	500-550	✓	✓	✗	✗	✓	firm	gentle slope	T69
	Sgurr Dhomhnuill SE	650	✓	✓	✓	✓	✓	firm	vegetated, few crags, mainly free face crags	T70
	Sgurr Dhomhnuill NE	550-610	✓	✓	✗	✓	✓	firm	topographic change	T71
	Meall Mor SE	680	✓	✓	✗	✓	✓	firm		T72
	Stob a' Chuir W	670	✓	✓	✗	✗	✓	firm		T73
	Stob a' Chuir SW	640-680	✓	✓	✗	✗	✓	firm	shattered bedrock	T74

Glen	Hillslope	Altitude m	Fresh glacial evidence below	No glacial evidence above	Boulder concentrations at trimline	No clear periglacial evidence below	Widespread periglacial evidence above	Clarity	Comments	Trimline no.
	Stob Mhic Bheathain SE	550-640	✓	✓	✓550-640	✓	✓	firm	closely foliated around trimline	T75
	Stob Mhic Bheathain SSE	620	✓	✓	✓	✓	✓	firm	no outcrops, relies on debris evidence	T76
Cona	Sgurr an Lubhair SE	525	✓	✓	×	✓	✓	firm	no outcrops	T77
	Sgurr an Lubhair SW	520-620	✓ up to 350m	✓	×	✓	✓	firm	vegetated, free face outcrops	T78
	Stob Mhic Bheathain N	640	✓	×	✓	✓	✓	firm		T79
	Meall nan Damh S	610	✓	✓	✓	✓	✓	firm	few outcrops	T80
	Meall nan Damh W	680	✓	✓	×	×	✓	firm		T81
	Meall Mor NE	700	✓	✓	✓	×	✓	very firm		T82
	Sgurr Ghiubhsachain SE	>/=670	✓	×	×	✓	✓	indistinct	topographic contol?, mixed evidence	T83
	Clac Garbh W	>/=700	✓	×	×	×	✓	indistinct	widespread scouring	T84
	Clac Garbh E	>/=670	✓	×	×	×	✓	indistinct	mixed evidence at all altitudes	T85

Glen	Hillslope	Altitude m	Fresh glacial evidence below	No glacial evidence above	Boulder concentrations at trimline	No clear periglacial evidence below	Widespread periglacial evidence above	Clarity	Comments	Trimline no.
Eil	Druim Fada SE	580-650	✓	✓	✓	✓	✓	firm	peat cover, few outcrops	T86
	Druim Fada S	575-620	✓	✓	✓	✓	✓	firm	peat cover, no outcrops	T87
	Stob Coire a' Chearcaill NE	550-600	✓	✓	×	×	✓	firm	peat, no outcrops	T88
	Sgurr Craob a' Chaorain N	660	✓	×	✓	✓	✓	firm		T89
	Sgurr Craob a' Chaorain NE	700	✓	×	×	✓	✓	firm		T90
Suileag	Druim Fada W	610	×	✓	×	✓	✓	firm		T91
	Meall Onfhaidh W	640	✓	✓	×	✓	✓	firm	few bedrock exposures, topographic change	T92
	Meall Onfhaidh E	630-650	✓	✓	✓	✓	✓	firm		T93
	Meall a' Phubuill SE	650	✓	✓	×	×	✓	firm	no outcrops	T94
	pt 747 W	690	✓	✓	×	✓	✓	firm		T95
	Druim Gleann Laoigh S	650	✓	✓	✓	✓	✓	firm	no outcrops	T96
	Druim Fada NE	610-630	×	✓	×	✓	✓	firm		T97
Fionnlaighe	Na h-Uamhachan SW	(>/=) 630	✓	×	✓	×	✓	firm		T98
	Sron Liath S	>700	✓	×	×	✓	×	probable	massive gneiss	T99
	Gulvain S	670-720	✓	✓	×	✓	✓	firm		T100

Glen	Hillslope	Altitude m	Fresh glacial evidence below	No glacial evidence above	Boulder concentrations at trimline	No clear periglacial evidence below	Widespread periglacial evidence above	Clarity	Comments	Trimline no.
	Meall Onfhaidh N	540-681	✗	✓	✗	✗	✓	indistinct	steep, few outcrops, shattered	T101
	Meall a' Phubuill N	700	✓	✓	✗	✗	✓	probable	steep. few outcrops, shattered	T102
Dubh Lighe	Beinn an Tuim E	670	✓	✗	✓	✓	✓	firm		T103
	Streap SE	660	✓	✓	✗	✗	✓	firm		T104
Finnan	Fraoch-Bheinn E	630-780	✓	✗	✗	✓	✓	probable	little clear periglacial evidence	T105
	Beinn an Tuim W	600-780	✓	✓	✓600-650	✓	✓	probable	steep slopes around trimline	T106
	Beinn an Tuim NW	650-770	✓	✓	✓650-700	✓	✓	probable		T107
	Sgurr a' Choire Riabhaich SE	(670-) 745	✓	✓	✗	✓	✓	firm		T108
	Sgurr Thuilm E	770-800	✓	✓	✗	✗	✓	firm	steep crags, vegetated slope	T109
	Streap W	760-800	✓	✓	✗	✗	✓	firm		T110
Ardnamurchan	Beinn Laga S	none						indistinct		T111

Table A.4.2 Glacial and periglacial evidence on low summits

Summit	Altitude m	Clear periglacial evidence	Clear, fresh glacial evidence	Perched blocks	Position with respect to adjacent trimlines	Summit no.
Creag Bhan	500	✓	✓	✓	above	S1
Seann Chruach	521	✓	×	✓	above	S2
Ceann Loch Uachdrach	328	✓	✓	✓	above	S3
Creag nan Lochan	498	(✓)	✓	×	above	S4
Garbh Coire	320	✓	(✓)	(✓)	above	S5
Meall Daimh	575	×	✓	✓	below	S6
Sgurr a' Bhuic	450	✓	×	✓	below	S7
Sgurr na Laire	620	✓	✓	✓	below?	S8
Sgorr nan Cearc	668	✓	×	✓	below	S9
A' Bheinn Bhan	477	×	✓	✓	below	S10
Druim Leathad nam Fias	559	✓	(✓)	✓	below	S11
Meall a' Choire Chruinn	634	✓	×	✓	below	S12
Glas Bheinn	635	✓	×	✓	below	S13
Sron Liath	765	(✓)	✓	✓	above	S14
Meall Onfhaidh	681	(✓)	×	✓	above	S15

Table A.4.3 Glacial and periglacial evidence on cols

Col	Altitude m	Clear periglacial evidence	Clear, fresh glacial evidence	Perched blocks	Position with respect to adjacent trimlines	Col no.
Head of Glenfinnan	471	✗	✓	✓	below	C1
Head of Glen Dubh Lighe	350	✗	✓	✓	below	C2
Head of Glen Fionnlaighe	470	✗	✓✓	✓	below	C3
Meall a' Phubuill W	380	✓	✓	✓	below	C4
Meall a' Phubuill E	650	✓	✓	✓✓	below	C5
Na h-Uamhachan N	580	✓	✓	✓	below	C6
Rois-bheinn N	550	✓✓	✓	✓	below	C7
Rois-bheinn E	700	✓	✗	✓	above	C8
An t-Slat-Bheinn E	340	✗	✓	✓	below	C9
Beinn Mhic Cedidh E	480	✗	✓	✓	below	C10
Beinn Mhic Cedidh W	350	✓	✓✓	✓✓✓	below	C11
Beinn Odhar Beag N	760	✓✓	✓	✓	above	C12
Beinn Odhar Beag S	690	✓✓	✓	✓	above	C13
Sgurr Ghiubhsachain E	580	✓	✓	✗	below	C14
Sgurr Ghiubhsachain WSW	710	✓✓	✗	✗	above	C15
Clac Garbh E	370	✓	✓✓	✓✓	below	C16
Meall Mor W	650	✓	✓	✓	below	C17
Meall Mor E	500	✓	✓✓	✓✓✓	below	C18
Carn na Nathrac S	400	✗	✓✓	✓✓	below	C19
Sgurr na h'Ighinn SE	490	✓✓	✓✓	✓✓	below	C20
Beinn na h'Uamha E	250	✗	✓	✓	below	C21
Beinn Bheag E	490	✓	✗	✗	below	C22
Garbh Bheinn N	590	✓	✓	✓✓	below	C23
Sgurr Mhic Eacharna E	170	✗	✓	✓	below	C24
Beinn Resipol E	400	✗	✓	✓	below	C25

Table A.4.4 Bedrock slab relief above and below trimlines

	Below trimlines			Above trimlines		
Hill	altitude	average relief	standard deviation	altitude	average relief	standard deviation
Sean Cruach N	200	5.56	2.98	500	18.44	13.03
	445	6.36	7.03			
Beinn Gaire N	440	3.5	2.77	640	15.03	10.3
Meall a' Bhragaid N	150	4.57	3.7	400	12.19	12.13
	350	5.11	3.69	450	15.25	13.42
Druim Fada S	185	2.74	2.3	680	7.25	3.44
	460	5.47	3.16	735	9.33	19.57
Beinn an Tuim W	440	5.14	3.95	810	13.81	10.59
	575	6.62	5.03			
An Stac E	150	8.39	5.6	770	20.22	14.95
	490	5.72	4.26			
Beinn Beag NE	410	3.51	1.98	590	9.19	5.91
	475	6.33	6.2	655	17.17	21.75
Stob Mhic Beathain SE	320	2.5	1.09	660	6.03	4.1
				715	25.9	19.28
Sgurr Dhomhnuil NE	450	3.06	2.28	630	7.69	11.39
SE	560	6.47	5.43			
Meall a'Chorein.... N	195	4.61	3.26	485	14.42	16.44
	310	3.83	2.43	530	13.66	13.24
	390	3.41	2.29			

Table A.4.5 Bedrock joint depths above and below trimlines

Hillslope no.	2	2	2	2	2	4	4	4	4	5	5	5	6	6	9	9	9	11
Altitude - m	200	445	460	500	500	390	470	525	665	350	670	740	350	710	610	700	870	280
Position with respect to trimline	below	below	below	above	above	below	below	at	above	below	at	above	below	above	below	below/at	above	below?
0-5cm	20	18	16	14	17	19	15	9	12	19	16	11	19	11	16	9	8	17
5.5-10cm	0	2	4	6	3	1	5	10	5	0	4	5	0	7	4	11	12	3
10.5-15cm	0	0	0	0	0	0	0	0	2	1	0	4	1	1	0	0	0	0
15.5-20cm	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
20.5-25cm	0	0	0	0	0	0	0	1	1	0	0	0	0	1	0	0	0	0
25.5-30cm	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
>30cm	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
mean	2.03	3.1	3.4	4.33	3.15	2.93	4.38	6.28	6.13	2.75	3.48	6.5	2.75	6.05	4.08	5.85	5.9	3.85
standard deviation	0.92	1.34	2.35	2.51	1.64	1.38	1.83	4.02	5.05	2.35	1.8	3.2	2.35	4.27	1.9	2.57	2.63	1.75
Hillslope no.	11	13	13	15	15	15	16	16	20	20	21	21	21	22	22	22	25	25
Altitude - m	460	400	460	280	605	713	440	640	630	773	760	882	882-slab	360	450	480	580	640
Position with respect to trimline	above?	below?	below?	below	above	above	below	above	below	above	?	above	above	below	below	at	below	below
0-5cm	17	15	13	17	15	10	18	19	12	8	16	7	20	16	17	13	17	16
5.5-10cm	3	5	7	3	5	9	2	1	8	9	3	4	0	3	3	5	3	3
10.5-15cm	0	0	0	0	0	0	0	0	0	3	1	1	0	1	0	0	0	1
15.5-20cm	0	0	0	0	0	0	0	0	0	0	0	3	0	0	0	2	0	0
20.5-25cm	0	0	0	0	0	0	0	0	0	0	0	1	0	0	0	0	0	0
25.5-30cm	0	0	0	0	0	0	0	0	0	0	0	3	0	0	0	0	0	0
>30cm	0	0	0	0	0	1	0	0	0	0	0	1	0	0	0	0	0	0
mean	3.58	4.33	5.03	3.43	4.35	8.3	3.15	3.13	4.63	5.9	4.23	14.35	2.45	3.8	3.45	6.28	3.05	3.8
standard deviation	1.69	1.87	2.11	2.05	1.64	11.09	1.35	1.11	2.1	3.35	2.29	14.41	1.02	2.3	1.57	4.25	1.86	2.44
Hillslope no.	25	25	27	27	28	28	28	28	28	28	30	30	30	30	30	32	32	33
Altitude - m	670	756	610	680	330	530	630-side	630-top	660	849	340	480	580-slab	580-cliff	756	540	650	400
Position with respect to trimline	at	above	below	above	below	below	below	below	at	above	below	below	below	below	above	below	above	below
0-5cm	13	9	7	10	17	17	16	8	9	8	17	11	19	10	9	11	16	18
5.5-10cm	6	6	6	4	1	2	4	8	7	6	3	6	1	5	6	6	3	2
10.5-15cm	0	2	1	1	2	1	0	4	1	0	0	2	0	3	2	2	1	0
15.5-20cm	1	1	1	0	0	0	0	0	0	0	0	1	0	1	1	1	0	0
20.5-25cm	0	1	3	1	0	0	0	0	0	0	0	0	0	0	1	0	0	0
25.5-30cm	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
>30cm	0	1	2	4	0	0	0	0	3	6	0	0	0	1	1	0	0	0
mean	5.7	10.43	11.48	13.48	4.3	3.9	3.33	6.9	14.58	19.38	3.1	6.43	2.4	7.5	10.43	6.25	4.38	3.25
standard deviation	3.08	11.25	10.19	16.38	2.86	3.04	1.79	3.54	26.3	22.99	2.14	4.88	1.34	8.33	11.25	3.8	2.42	2.13

Hillslope no.	33	33	34	34	34	37	37	37	37	38	38	38	38	38	39	39	39	41
Altitude - m	610	756	400	680	790	390	425	460	500	320	440-slope	440-top	440-cliff	702	510	660	702	500
Position with respect to trimline	below	above	below	above	above	below	below	at	above	below	below	below	below	above	below	above?	above	below
0-5cm	15	5	18	8	11	14	11	9	14	18	16	12	5	14	15	12	14	13
5.5-10cm	4	12	2	7	3	6	7	7	2	2	3	5	6	3	4	6	3	5
10.5-15cm	1	0	0	1	2	0	1	4	4	0	1	2	1	1	1	2	1	2
15.5-20cm	0	0	0	3	0	0	1	0	0	0	0	1	2	0	0	0	0	0
20.5-25cm	0	0	0	0	0	0	0	0	0	0	0	0	3	0	0	0	0	0
25.5-30cm	0	1	0	0	1	0	0	0	0	0	0	0	0	2	0	0	2	0
>30cm	0	2	0	1	3	0	0	0	0	0	0	0	3	0	0	0	0	0
mean	4.75	12.38	3.25	9.13	14.08	4.13	5.98	7.13	5.71	2.79	3.65	5.88	13.7	6.43	3.48	5.45	6.43	4.83
standard deviation	3.16	14.68	2.13	7.92	21.63	1.63	3.83	3.6	3.49	1.94	2.38	3.98	15.31	7.72	2.95	2.97	7.72	3.28
Hillslope no.	41	41	42	42	42	42	43	43	43	44	44	44	45	45	45	46	46	46
Altitude - m	610	720	500	620	640	740	440	580	740	300	330	490	270-slab	270	687	380	490	687
Position with respect to trimline	below?	above	below	at?	above?	above	below	below	above	below	below	at	below	below	above	below	at?	above
0-5cm	14	12	13	17	13	10	18	14	10	17	9	11	20	12	12	19	14	12
5.5-10cm	4	7	5	2	3	4	2	4	4	2	10	8	0	7	7	1	5	7
10.5-15cm	1	1	2	1	3	2	0	2	2	1	1	1	0	1	1	0	1	1
15.5-20cm	0	0	0	0	0	3	0	0	3	0	0	0	0	0	0	0	0	0
20.5-25cm	0	0	0	0	1	1	0	0	1	0	0	0	0	0	0	0	0	0
25.5-30cm	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
>30cm	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
mean	5.05	4.9	4.83	3.55	5.58	7.78	2.95	4.83	7.78	3.95	5.68	5.2	1.98	5.78	5.1	2.98	4.63	5.1
standard deviation	5.89	2.77	3.28	3.22	5.62	6.47	1.65	3.14	6.47	1.93	2.03	2.23	0.6	2.8	2.55	1.41	3.03	2.55
Hillslope no.	50	50	50	50	50	50	50	50	50	50	50	50	50	51	51	51	51	51
Altitude - m	100	150	200	200	250	300	350	370	370	370-felsite	400	435	450	200	400	600	800	885
Position with respect to trimline	below	below	below	below	below	below	below	at	at	at	above	above	above	below?	above	above	above	above
0-5cm	19	20	14	19	20	19	20	14	20	11	18	14	19	10	10	7	7	4
5.5-10cm	1	0	6	1	0	1	0	4	0	5	2	3	1	8	7	10	8	5
10.5-15cm	0	0	0	0	0	0	0	2	0	1	0	3	0	0	0	0	1	2
15.5-20cm	0	0	0	0	0	0	0	0	0	1	0	0	0	0	2	2	0	0
20.5-25cm	0	0	0	0	0	0	0	0	0	1	0	0	0	0	0	0	0	1
25.5-30cm	0	0	0	0	0	0	0	0	0	1	0	0	0	1	1	1	1	0
>30cm	0	0	0	0	0	0	0	0	0	0	0	0	0	1	0	0	3	8
mean	2.48	1.85	4.3	2.78	1.88	2.55	1.75	5.3	2.4	7.75	3.23	5.03	3.23	7.83	7.63	7.9	13.1	22.55
standard deviation	1.4	0.71	2.62	1.16	1.05	2.52	0.7	3.06	0.84	6.94	1.9	2.99	1.34	8.02	6.62	5.04	14.42	20.75

Hillslope no.	52	52	52	53	53	53	53	55	55	55	55	55	59	59	59	59	60	60
Altitude - m	110	200	300	235	360-slab	360-cliff	430	330	450	540	590	680	210	345	410-slab	410	280	490
Position with respect to trimline	?	above?	above	below	below	below	above	below	below	below	above	above	below	below	above	above	below	below
0-5cm	19	14	17	19	19	9	13	15	18	8	4	5	18	16	20	10	18	9
5.5-10cm	1	5	3	1	1	9	6	3	2	7	10	6	2	4	0	7	2	8
10.5-15cm	0	0	0	0	0	1	0	2	0	2	3	2	0	0	0	1	0	1
15.5-20cm	0	1	0	0	0	1	1	0	0	1	2	2	0	0	0	0	0	1
20.5-25cm	0	0	0	0	0	0	0	0	0	0	1	1	0	0	0	0	0	0
25.5-30cm	0	0	0	0	0	0	0	0	0	0	0	1	0	0	0	2	0	1
>30cm	0	0	0	0	0	0	0	0	0	2	0	3	0	0	0	0	0	0
mean	2.71	5.43	3.43	2.6	3.05	6.4	5.65	4.4	3.2	14.83	9.08	16.48	3.13	3.6	2.98	7.8	3.29	6.78
standard deviation	1.02	3.56	1.59	1.18	1.09	3.62	4.04	3.55	2.08	24.87	4.91	16.54	2.14	2.16	0.64	7.11	1.99	6.03
Hillslope no.	60	60	61	61	61	62	62	68	68	68	69	69	70	70	71	71	73	73
Altitude - m	590	680	420	510	560	420	560	410-cliff	420	598	400	460	410	550	400	510	560	720
Position with respect to trimline	above	above	below	?	above?	below	above?	below	below	above	below	below	below	below	below	below	below	above
0-5cm	9	5	19	13	16	18	16	3	15	5	16	8	13	7	15	11	16	8
5.5-10cm	7	6	0	6	3	2	3	7	4	10	4	10	4	6	2	1	4	8
10.5-15cm	3	2	1	0	1	0	1	5	1	2	0	2	1	5	2	2	0	2
15.5-20cm	0	2	0	1	0	0	0	0	0	2	0	0	1	2	0	2	0	1
20.5-25cm	0	1	0	0	0	0	0	1	0	0	0	0	0	0	0	0	0	0
25.5-30cm	1	1	0	0	0	0	0	0	0	1	0	0	1	0	1	0	0	0
>30cm	0	3	0	0	0	0	0	4	0	0	0	0	0	0	0	4	0	1
mean	7.53	16.48	3.08	5.03	4	3.03	4	15.95	4.15	9.13	4.15	6.15	6.55	7.7	5.15	14.53	3.5	7.85
standard deviation	6.4	16.54	2.26	3.38	2.22	1.26	2.22	14.55	2.32	5.77	2.07	3.11	6.21	5.15	5.96	17.14	1.76	7
Hillslope no.	74	74	74	74	75	75	75	75	77	77	79	79	82	82	83	83	83	84
Altitude - m	440	500	680	720	460	470	550	721	470	720	600	721	430	570	520	610	849	700
Position with respect to trimline	below	below	at?	above	below	below	below	above	below	above	below	above	below	below	below	below	above	?
0-5cm	18	8	8	8	18	15	10	10	13	5	17	10	17	11	16	7	8	7
5.5-10cm	1	7	7	8	1	3	7	4	7	6	2	4	2	8	3	9	6	11
10.5-15cm	1	3	4	2	0	0	2	1	0	3	0	1	1	1	0	2	0	1
15.5-20cm	0	1	0	1	1	0	1	1	0	3	0	1	0	0	0	2	0	1
20.5-25cm	0	0	0	0	0	0	0	1	0	1	1	1	0	0	0	0	0	0
25.5-30cm	0	0	0	0	0	1	0	2	0	1	0	2	0	0	0	0	0	0
>30cm	0	1	1	1	0	1	0	1	0	1	0	1	0	0	0	0	6	0
mean	3.6	7.98	7.8	7.85	4.35	6.35	6.45	12.95	4.8	11.03	4.33	12.95	3.3	5.03	5.5	7.3	19.38	6.78
standard deviation	1.96	7.28	7.44	7	4.41	8.14	4.84	19.6	1.98	8.47	4.87	19.6	2.5	3.02	5.88	4.11	22.99	3.59

Hillslope no.	85	85	85	95	98	98	98	99	101	101	103	103	103	104	104	105	105	105
Altitude - m	490-slab	490	630	710	590	680	691	690	430	610	510	670	810	620	909	630	760	790
Position with respect to trimline	below	below	below	above	below	above	above	below?	below	below?	below	at	above	below	above	below	above?	above
0-5cm	20	12	10	6	5	20	8	11	14	5	18	13	9	15	9	20	16	12
5.5-10cm	0	6	7	4	6	0	6	6	5	11	2	7	7	2	5	0	3	6
10.5-15cm	0	0	1	4	2	0	1	1	1	3	0	0	1	1	2	0	1	1
15.5-20cm	0	2	1	1	3	0	1	0	0	1	0	0	0	0	0	0	0	0
20.5-25cm	0	0	0	1	1	0	1	1	0	0	0	0	2	1	0	0	0	1
25.5-30cm	0	0	1	0	0	0	3	0	0	0	0	0	0	0	1	0	0	0
>30cm	0	0	0	4	2	0	0	1	0	0	0	0	1	1	3	0	0	0
mean	2.38	6.18	6.73	19.82	14.1	2.5	10.19	8.95	4.1	7.98	3.72	4.43	8.7	6.45	14.63	1.85	4.5	5.85
standard deviation	0.92	3.84	6.3	23.47	14.61	1.03	9.09	15.09	2.54	4	1.69	1.69	1.79	7.83	7.67	20.71	0.4	2.42

Hillslope no.	108	108	108	109	109	109	C23	S11	S11	S11	Opposite T70	Opposite T70	Opposite T71	South of T93	South of T93			
Altitude - m	670	720	852	770	840	963	590	360	430	550	330	420	420	480	550			
Position with respect to trimline	below	below	above	at/below	above	above	below	below	below	below	below	below	below	below	below			
0-5cm	20	16	13	19	12	14	8	16	13	8	18	11	13	11	12			
5.5-10cm	0	4	6	1	8	2	7	4	6	5	2	7	4	4	6			
10.5-15cm	0	0	0	0	0	1	0	0	1	3	0	1	3	4	3			
15.5-20cm	0	0	0	0	0	2	1	0	0	1	0	0	0	0	0			
20.5-25cm	0	0	1	0	0	1	1	0	0	2	0	0	0	0	0			
25.5-30cm	0	0	0	0	0	0	2	0	0	1	0	1	0	1	0			
>30cm	0	0	0	0	0	0	1	0	0	0	0	0	1	0	0			
mean	2.4	4.2	5.5	2.7	4.78	6.75	14.25	3.98	5	9.1	3.08	6.86	7	6.23	2.9			
standard deviation	5.09	0.75	1.77	4.68	1.21	2.31	23.27	2.42	3.09	5.58	1.76	7.32	11.55	5.84	3.95			

Appendix 5 Stratigraphic evidence from enclosed depositional basins

Table A.5.1 Basins investigated from which cores were not retrieved for pollen analysis.

Trough	Location	Grid Reference	Depth - cm	Basal sediments	Comments
Linnhe	Glac Daraich - kettlehole	NM 003 639	630	peat above sands and gravels	no early postglacial limnic sediments
Linnhe/ Tarbert	Lochan Torr an Fhuich	NM 925 583	650	green gyttja above silty clay	no interstadial sediments
Shiel	Acharacle ice contact slope	NM 674 686	511	peat above sands and gravels	no early postglacial sediments
Shiel/ Moidart	Lochan nan Chriochan	NM 721 705	484	organic muds	no interstadial sediments
Shiel/ Moidart	Captain Robertson's Cairn	NM 722 708	638	peat with macro fossils	no interstadial sediments
Shiel/ Moidart	Lochan Bad na Sgitheiche, south	NM 714 706	386	amorphous peat with macro fossils	no interstadial sediments
Shiel/ Moidart	Lochan Bad na Sgitheiche, edge	NM 714 707	434	amorphous peat	no interstadial sediments

A.5.2

Pollen Chronology of sediments near Acharacle

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Methods

The basal organic sediments of two basins near the mouth of the Shiel trough were pollen analysed to determine the ages of these sediment accumulations. The geomorphic setting of these basins is fully described in Section 5.3.2. The large basin of BNBio was depth-probed using an Eijelkamp narrow-diameter (2.5cm) gouge and extension rods on a simple line of traverses through the topographic centre of the basin. At the deepest point located, the basal 25.0cm of sediment were retrieved by a narrow-diameter (2.5cm) Abbey piston corer. The basin BNBio/2 is a very much smaller basin, only a few tens of square metres in extent. The site sampled lies in the centre of the basin. A 50.0cm long Abbey core (diameter 5.0cm) was obtained from the basal organic sediments. In Figure 2 the depths are presented as depths within the core sampled; the depths below the ground surface are 235-285cm. The basal 4cm of the retrieved core contain minerogenic sandy silts, and below this are a further 95cm of similarly inorganic sediment with no suggestion of organic input. Following extrusion in the field (compression c.10-12% in both cases) into clean drain pipes and being wrapped in plastic, the cores were stored at $4^{o}C$.
Samples of 0.25cm thickness were prepared for analysis by standard chemical techniques (Moore, Webb & Collinson 1991), including 10µm sieving and hot hydrofluoric acid treatment to remove silicates. Measures of pollen concentration were not obtained. Residues were stained with safranin and embedded in silicon oil. Analyses were made on an Olympus BX 40 microscope, usually at mag. x400. Samples from BNBio at 572.5cm, 577.5cm and 580.0cm had very few pollen grains (Fig 1b; grains per traverse) and were counted at mag. x200, grains being checked at mag. x400. The purpose of the analysis was simply to establish a chronology for the basal sediments, and so the sum counted was set at 100 total land pollen (tlp); *Alnus - Thalictrum* inc. on Figures 1a and 2a. In the core from BNBio the sample at 580.0cm was barely polleniferous (that at 575.0cm proved uncountable) and counts are very low, but this does not affect the chronological interpretation.
In addition to conventional percentage-based analyses, pollen preservation was determined according to Tipping (1987a), and microscopic charcoal fragments and sulphide spherules (Wiltshire, Edwards & Bond 1992) recorded.

Results

In both basins the basal organic sediments almost certainly date to the earliest postglacial (Holocene). The pollen chronology for the first c. 1000 years following the Loch Lomond Stadial is very well established for western Scotland (Lowe & Walker 1981, 1991: Walker & Lowe 1985, 1987; Tipping 1988, 1989a; Lowe & Cairns 1991; Benn, Lowe & Walker 1992). Its consistent 'successional' sequence has regional significance, and can be applied to the region close to Loch Shiel also (Williams 1977; Wain-Hobson 1981; Shennan et al. 1993, 1994).
This temporal pattern can be recognised in the sequences analysed at BNBio and BNBio/2. The pattern is not distorted by differential pollen preservation, well-preserved grains representing a clear majority of the total (Figs 1b, 2b).
At BNBio the earliest pollen spectra (587.5-582.5cm) depict a landscape devoid of trees, with *Juniperus* pollen relatively common. This phase cannot be dated precisely since the peak in *Juniperus* pollen is not synchronous across the British Isles (Tipping 1987b), but probably pre-dates c.9800 - $9600^{14}C$ BP, the first palynological appearance of *Betula*. This horizon occurs at c.580.0cm, and the final spectrum at 570.0cm probably records the regional colonisation of *Corylus/Myrica,* dated to $c.9200^{14}C$ BP at the majority of sites (Boyd& Dickson 1986; Birks 1989). Pollen assemblages earlier in the 'succession' than the expansion of *Juniperus,* such as the Graminae-*Rumex* and *Empetrum* phases known from other sites, are not recorded at BNBio. *Empetrum* is present, but at lower proportions than Ericaceae, possibly representing pedological controls on the quartzitic rock (Tipping 1989b). The absence of 'pioneer' assemblages rich in *Rumex* and bare ground herbs such as *Artemisia*

may have no chronological or glaciological significance (cf. Lowe & Walker 1981; Benn et al. 1992) given the comparatively crude defining of the deepest point in the basin (above) and the absence of pollen analyses from the basal c.3cm of sediment (Tipping 1988).

At BNBio/2 a similar pattern can be discerned, though less clearly. A 'pionner' phase of Graminae-*Rumex* is more distinct at this site, but is complicated by the presence of *Juniperis* pollen at peak, though relatively subdued, values, and by the occurrence of successionally later *Betula*. Some contamination during sampling may be responsible for this, or these pollen grains may derive from sources distant from the basin. The anticipated subsequent peak in *Juniperus* (cf BNBio) is absent, possibly missed through the sampling interval being too wide in a comparatively compressed pollen stratigraphy. Ericaceae percentages expand above the Graminae-*Rumex* phase, and the limited representation of *Empetrum* in this area is confirmed by these analyses. *Betula* and *Corylus/Myrica* values seemingly rise together, again possibly a product of the compressed pollen stratigraphy and poor temporal resolution, and the uppermost count confirms a Holocene age for these sediments in the appearance of *Quercus, Ulmus* and *Alnus*.
In comparing the results from these two basins, there is no reason to assume significance in the apparent absence of 'pioneer' phases at BNBio. This feature need not imply the later deglaciation of BNBio (above; cf. Benn et al. 1992). Nvertheless, in both basins the earliest organic sediments to accumulate are clearly of earliest postglacial (Holocen2) age. At BNBio/2 these sediments are underlain by 99cm of minerogenic sediments, and their accumulation might be linked to sediment deposition during the Loch Lomond Stadial (see Section 5.3.2). One uncertainty relates to the comparatively small sediment thicknesss in the basin (several metres minerogenic sediment accumulated in basins dammed by Loch Lomond Stadial glaciers in the Cowal (Tipping 1986) and the Awe valley (Tipping 1989a)). However, at BNBio\2 the steep gradient of the valley may have led to drainage back into the ice or laterally, rather than into the basin.

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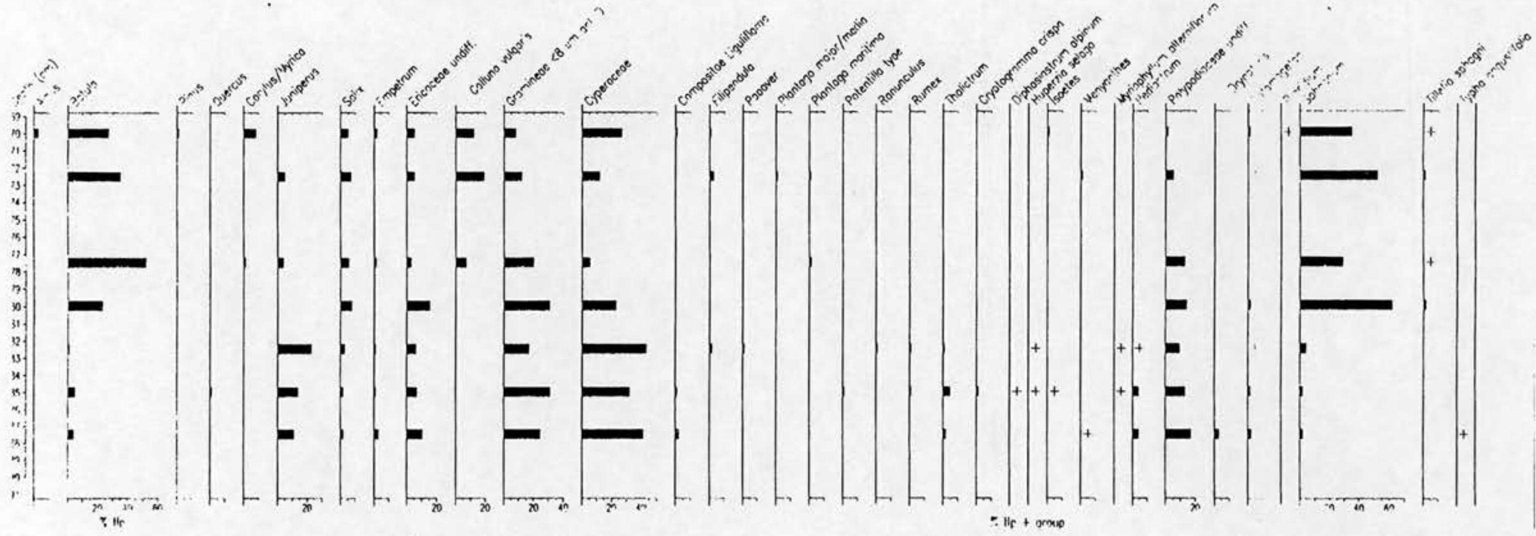


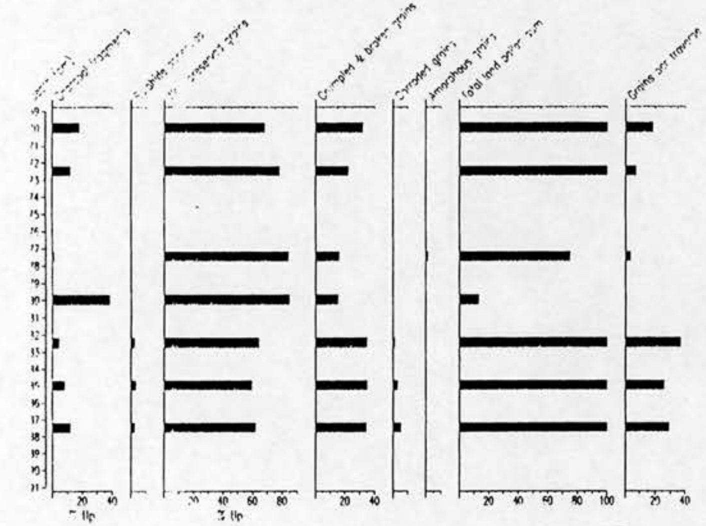
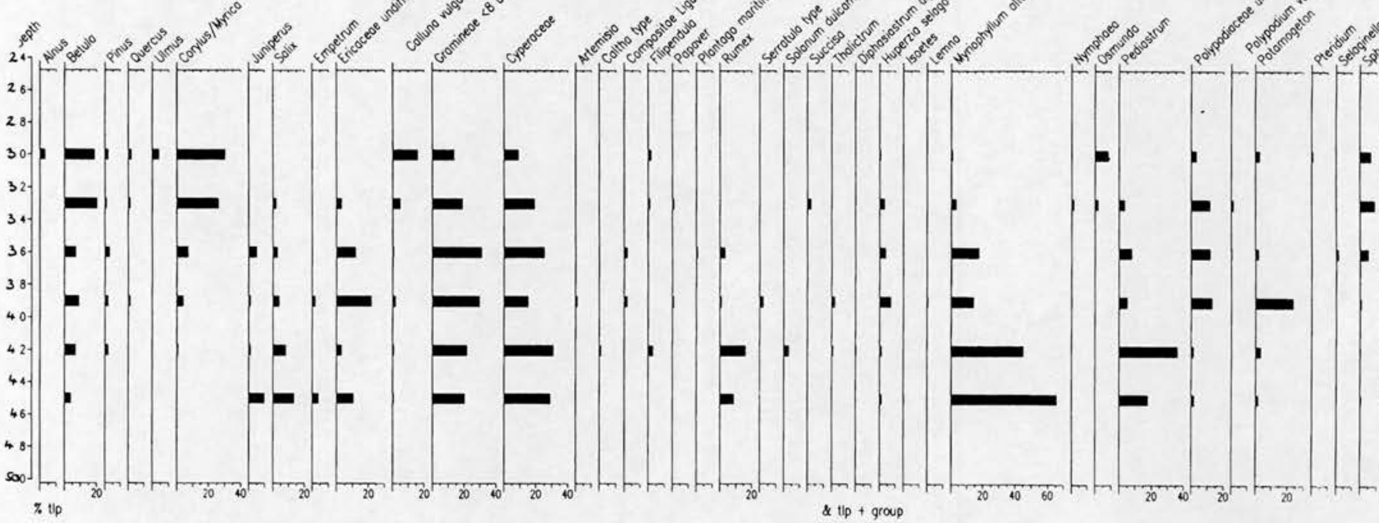


Fig 1. Pollen analyses of the basal 22.0cm of sediment at BNBio :- a) percentage-based analyses of all pollen taxa; b) percentage-based analyses of microscopic charcoal, sulphide spherules and pollen preservation, plus the tlp sum attained and a measure of pollen abundance, grains per traverse.



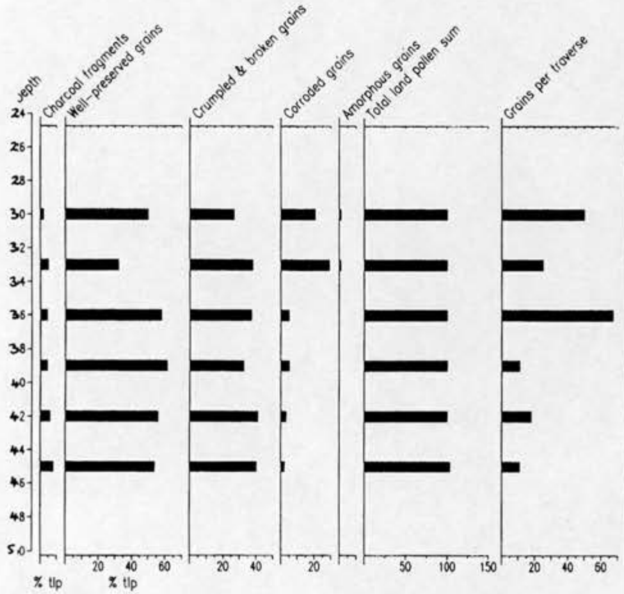
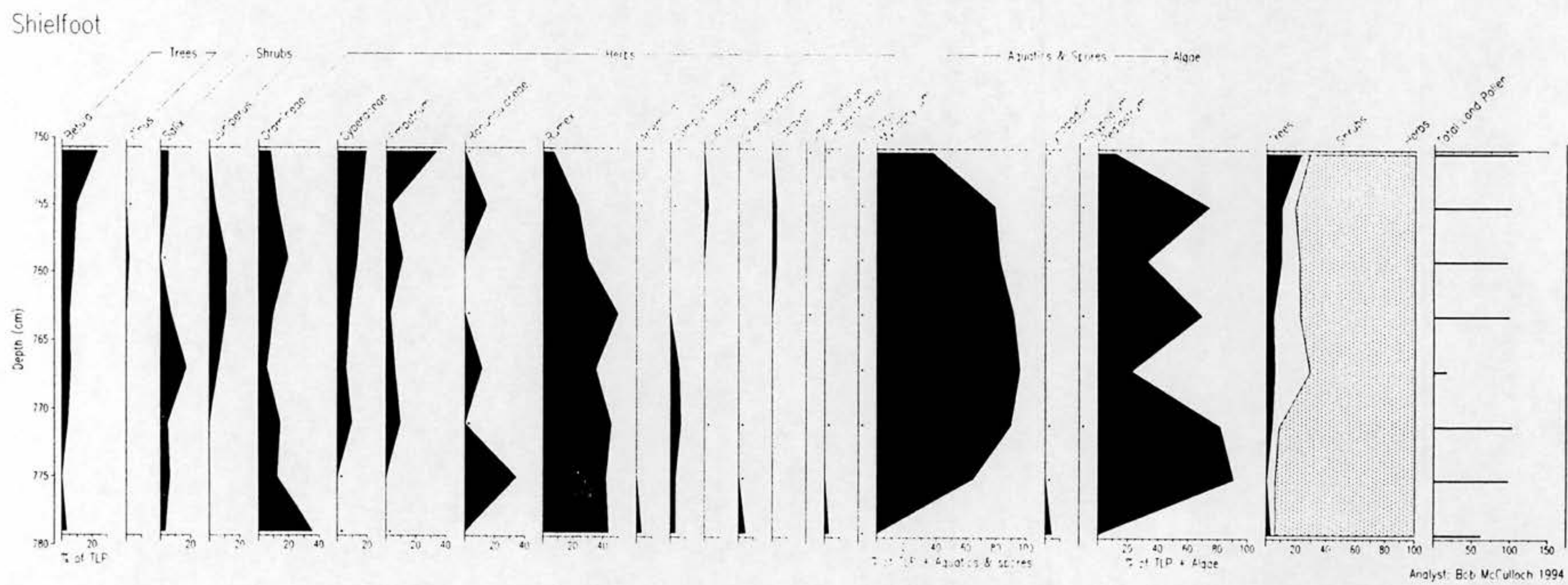


Fig 2. Pollen analyses of the basal 22.0cm of sediment at BNBio/2 :- a) percentage-based analyses of all pollen taxa; b) percentage-based analyses of microscopic charcoal, sulphide spherules and pollen preservation, plus the tlp sum attained and a measure of pollen abundance, grains per traverse.

A.5.3 Pollen analyses of the basal 30.0cm of sediment at Glac Mhor. **McCulloch, 1994.**



Appendix 6 Radiocarbon date

DESCRIPTION OF MATERIAL: Peat for bulk radiocarbon dating. Samples from top and bottom of peat raft.

POSITION IN STRATIGRAPHY: The raft of peat is near the base of a unit of clast supported, sub rounded and weakly stratified gravels and cobbles. This unit overlies blue/grey laminated and well sorted clay. These are interpreted as a fluvioglacial outwash plain overlying a distal glacimarine deposit.

POSSIBLE CONTAMINATION: None obvious

LOCATION: 5° 21.9'W 56°41.0'N
The peat raft is at 4.5m O.D. Newlyn in a river bank section. The section is cut into the lowermost of a series of river terraces incised into a floodplain that is graded to a sea level of approximately 10m O.D.

OTHER RELEVANT DATES: No local dates.

Result

DATE No.: AA - 12926
SAMPLE No.: IVS 93 - 5 peat
FRACTION OF MODERN CARBON: 0.6818 ± 0.0058
RADIOCARBON AGE: 3,075 ± 70 years before present (conventional radiocarbon age)

Appendix 7 Calculated calving fluxes

Table A.7.1 Calculated calving fluxes and truncated ablation areas for former tidewater glaciers in Western Lochaber.

Glacier	Width - m	*Mean water depth at terminus - m	Estimated ice cliff height - m	Mean glacier thickness - m	Calculated calving speed - ma^{-1}	Calculated calving flux $-m^3a^{-1}$	Equivalent extra ablation area necessary - km^2 #	Percentage of glacier area
Ailort	2350	15.2	20	35.2	197	$1.63x10^7$	5.4	7%
Moidart	700	8.9	15	23.9	144	$2.41x10^6$	0.8	2%
Sunart	950	10.8	20	30.8	160	$4.68x10^6$	1.6	2%
Linnhe	4100	59.5	50	109.5	566	$2.54x10^8$	85	15%

* = corrected for higher LLS sea-levels.
# = using an ablation rate at MSL of $3ma^{-1}$, derived from mass balance curves presented in Kerr (1992), on the basis of an ELA depression estimated from those of non-calving glaciers in Western Lochaber.

Fig 2.1 Glacial Geomorphology of Western Lochaber, Scotland